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Tropical Cyclone Forecasters Reference Guide
2. Tropical Climatology

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ABSTRACT

One of the keys to safe and successful naval operations in the tropics is a thorough understanding of tropical meteorology. The Tropical Cyclone Forecasters Reference Guide is designed primarily as a ready reference for mid-latitude forecasters required to provide tropical meteorology support to staff commanders. This technical note provides a comprehensive overview of tropical climatology and is chapter 2 of the reference guide. Subjects discussed include major factors which effect the tropical climate, tropical synoptic models, wave disturbances, tropical vortices and tropical cyclones.

ACKNOWLEDGMENTS

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The tropics are loosely defined as the area between 30° north and south of the equator. As Nieuwolt (1977) pointed out, the tropics constitute about 40% of the land surface of the earth and are inhabited by nearly 40% of the world population.

This section summarizes the climatology of the tropics. Knowledge of climatology may benefit the forecaster by providing information on what mean weather systems to expect in day-to-day forecasting situations (Landsberg, 1941). Using this information, along with quality data, such as satellite imagery, surface charts, and numerical forecast aids, a meteorologist will in time, develop forecast skill. The major purpose of this section is to reduce the on-site time required to become a proficient forecaster.

1. MAJOR FACTORS AFFECTING THE TROPICAL CLIMATE

1.1 The Earth's Heat-energy Balance

The primary energy source for meteorological and oceanic processes comes from the sun. Observations conducted over a period of years indicate that the solar radiation received by the earth is not changing appreciably from year to year and varies only with latitude and season.

The solar radiation falling perpendicularly upon a flat surface at the outer edge of the atmosphere is about 1376 watts per square meter (Hichey et al., 1980). This equates to 1.97 calorie per minute per square centimeter. Some of this energy is reflected or scattered back to space and plays no further role in the earth's heat-energy balance. The remainder is absorbed by the earth's surface and the atmosphere. Attenuation of solar radiation energy is greatest at high latitudes where the path through the atmosphere is longest. Also, solar energy reaching the surface at high latitude is spread over a larger area because the surface is at an angle to the incoming radiation. Therefore the earth's surface at the poles is a heat sink when compared with the tropics.

The outgoing energy, which is radiated to space by the earth's surface and atmosphere, is more intense in lower vice higher latitudes. Most of the energy that radiates from the atmosphere comes from the uppermost layers of water vapor in the atmosphere. Convective mixing of the water vapor in the tropics results in a very warm/deep/moist layer which radiates a moderate amount of the total energy received back into space. The result of this considerable incoming radiation and moderate outgoing energy is an over abundance of heat in low latitudes. Therefore, there must be a poleward heat transfer from the equator to prevent accumulation of excess heat. Hence, the tropics are a source region of heat for both the oceanic and atmospheric circulations.

Figure 2.1 (Sellers,1965) shows the average latitudinal distribution of the global radiation balance. The upper curve shows that the earth's surface receives a surplus of radiative energy at all latitudes except near the poles, therefore the earth's surface is a heat source. On the other hand, the atmosphere loses energy faster than it receives it, thus it is a heat sink. The sum of these two, (the energy balance of the earth-atmosphere system) shows that the net radiative energy in the atmosphere is positive in tropical belt and negative for in the polar regions. The figure also reveals that a two-mode heat transports exists. The earth's surface radiates energy to the atmosphere and the atmosphere transfers it from the equator to the poles. There is still considerable doubt as to the magnitudes of the radiation rates, but with the development of new sensors a more reliable earth radiation budget will eventually be computed.

1.2 Tropical General Circulation

The general tropical and subtropical circulation is usually described in terms of a set of statistical parameters of the entire atmosphere. These statistics can be displayed using numerous types of maps and graphs. A basic understanding of these statistics is the cornerstone in understanding the basic climatology of the tropical regions.

The climatology charts selected to define the general circulation, or statistical mean circulation, in the tropics are divided into three basins: Pacific Ocean, Indian Ocean, and Atlantic Ocean. For the first two basins, four seasons are discussed using monthly charts. The months chosen to represent the four seasons are different for the Pacific and Indian

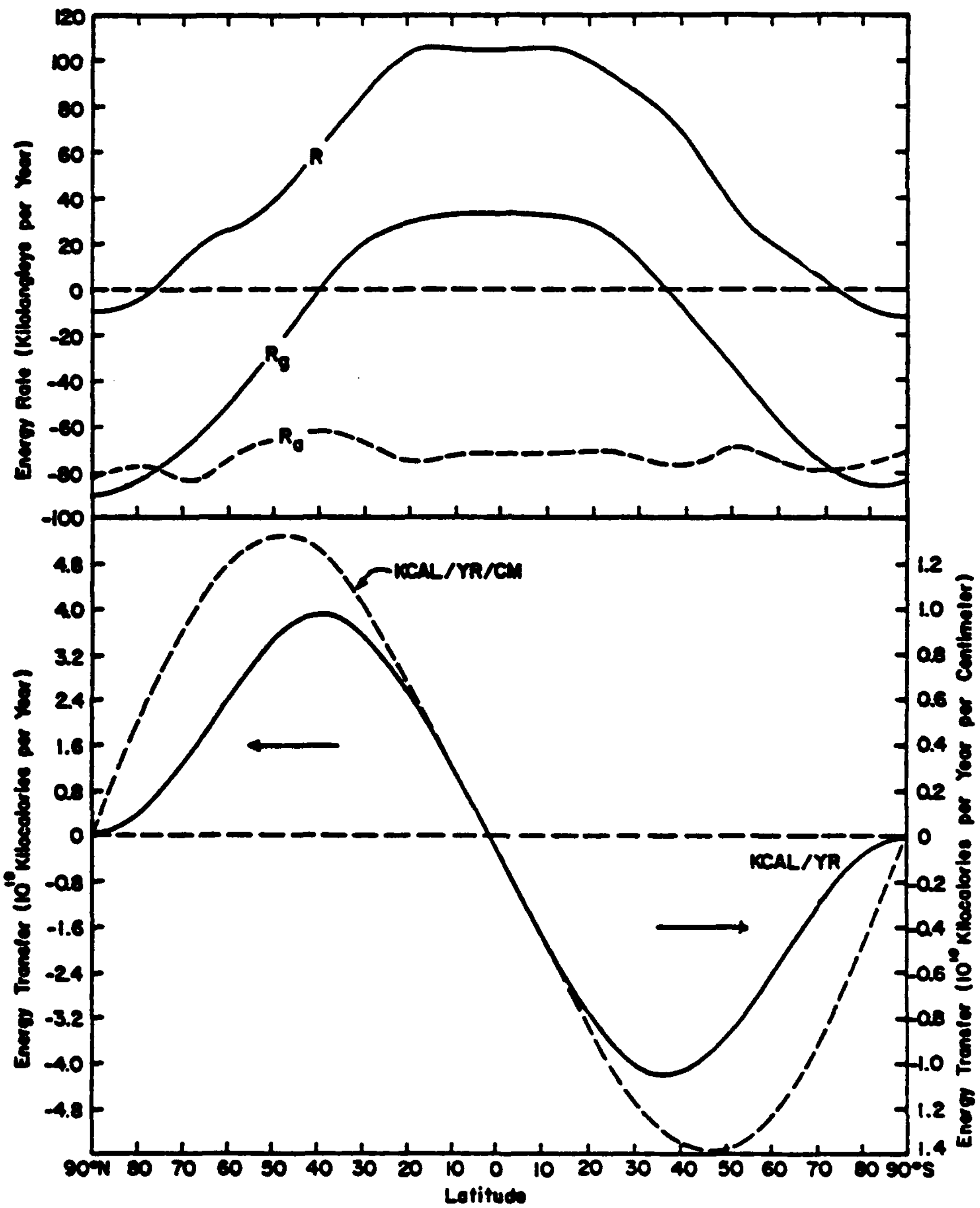

Fig. 2.1. The average annual latitudinal distribution of radiation balances of the earth's surface, R; of the atmosphere, Ra; and of the earth-atmosphere system, Rg; in kilolangleys per year (top) and the poleward energy flux in kilocalories per year and in kilocalories per year per centimeter of latitude circle (bottom). (Sellers, 1965).

Oceans due to the difference in the large scale synoptic patterns and timing of the transition seasons in these basins. For the Atlantic Ocean, only two months are discussed since the flow field in this basin changes very little during the year.

1.2.1 Pacific Ocean Surface Circulation

1.2.1.1 February

The winter subtropical ridge in the Northern Hemisphere is located at 25-30N (Fig. 2.2). Equatorward of the ridge, the northeasterly trade winds dominate the Northern Hemisphere tropics. The general circulation in the eastern and western halves of the Pacific are markedly different.

In the western half, the northeasterly trades cross the equator, become northwesterlies, and converge into the Intertropical Convergence Zone (ITCZ)/monsoon trough near 12S. This flow is known as the Australian summer monsoon.

East of the dateline, however, the cross equatorial flow is from the south and the ITCZ is located near 5N. This section of the ITCZ is actually only a trade wind confluent zone and very little convective activity takes place during this time of the year. The other feature of note is the Southern Pacific Convergence Zone (SPCZ) which lies near 20S from 175E to 160W. While the SPCZ is often viewed as an extension of the Australian monsoon trough, they are separate features and are not always connected.

1.2.1.2 May

In May the major synoptic features shift northward from the February positions (Fig. 2.3). The Northern Hemisphere (NH) subtropical ridge has moved north of 30N and the ITCZ in the eastern North Pacific is now located near 10N. The ITCZ has also changed its character. From Central America to the col at 110W, the ITCZ is more active as the northerly and southerly flows begin to oppose each other. It is in this region where most cyclogenesis takes place. West of the col, the ITCZ becomes a confluence zone with decreased convective activity.

In the western Pacific, the Southern Hemisphere ITCZ/near equatorial trough/ buffer zone and SPCZ have moved equatorward. The western North Pacific ITCZ/near equatorial trough or buffer zone begins to appear at 4N near the Caroline Islands (approximately 150E). Twin cyclones, one in the Northern and one in the Southern Hemisphere, are common at this time of year.

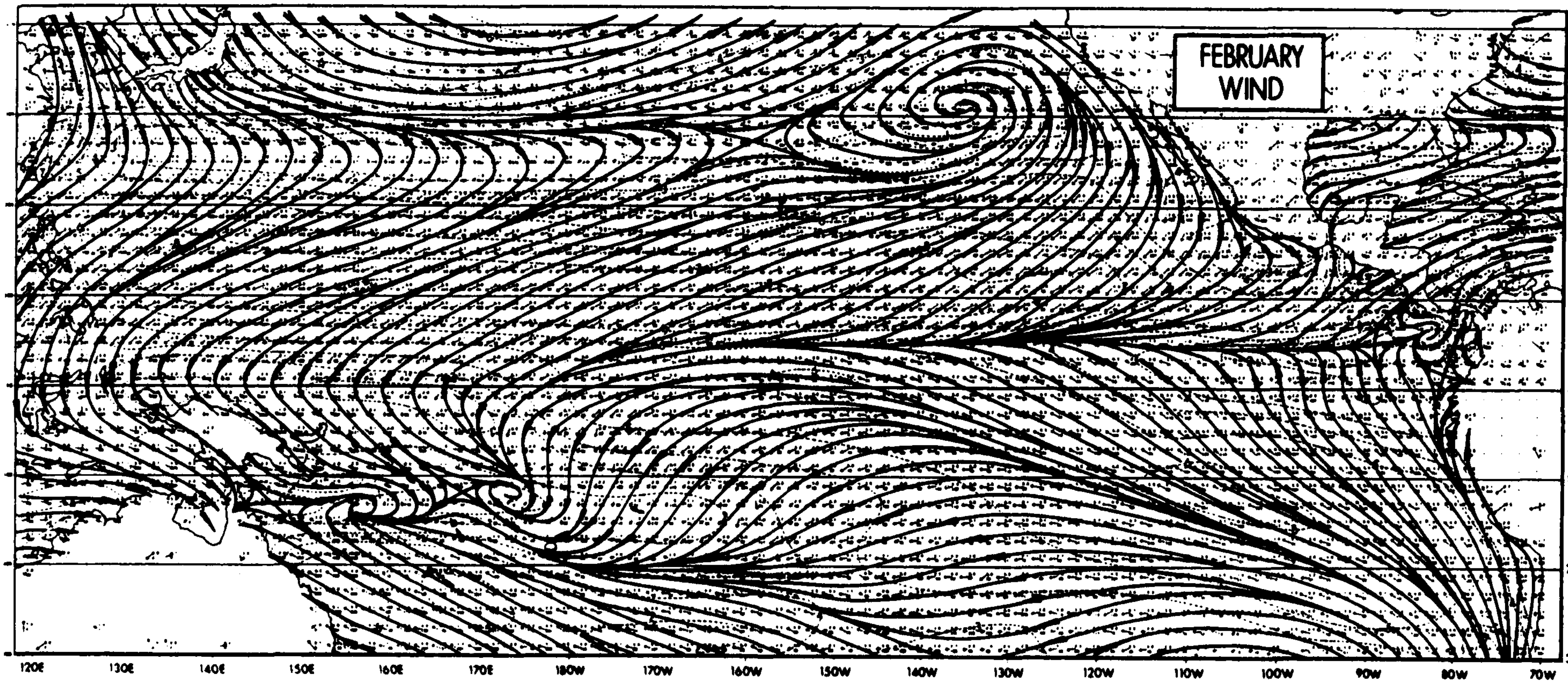

Fig. 2.2. Mean surface level streamline analyses over the Pacific for February (Sadler, 1975).

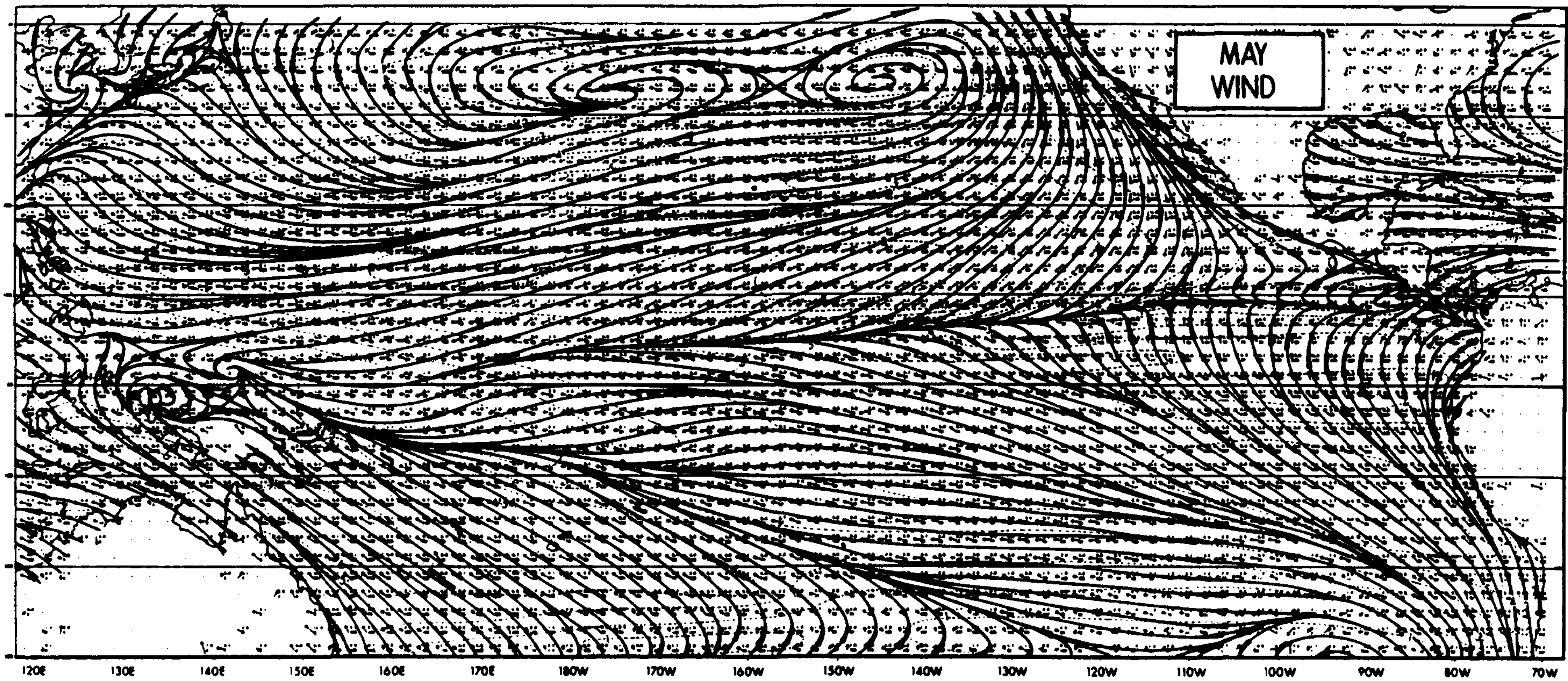

Fig. 2.3. Mean surface level streamline analyses over the Pacific for May (Sadler, 1975).

1.2.1.3 August

By August, the NH subtropical ridge has migrated to its northernmost latitude near 40N (Fig. 2.4). Cross equatorial flow from the Southern Hemisphere extends across the entire Pacific, converging into a nearly continuous ITCZ/Monsoon trough in the Northern Hemisphere.

The ITCZ in the eastern North Pacific has now moved north of 10N. The active portion of the ITCZ now extends west to 125W. The confluent portion of the ITCZ extends farther west to the point where it is nearly connected with the monsoon trough in the western North Pacific.

The monsoon trough of the western North Pacific is poorly depicted in a mean chart because it is often migratory. In general, it typically displays a southeast-northwest orientation. However, there are instances where the eastern portion of the trough migrates northward, reversing the orientation. Tropical cyclogenesis in the western North Pacific is largely confined to the monsoon trough, although other synoptic features such as the ITCZ/ near equatorial trough, fronts and the Tropical Upper Tropospheric Trough (TUTT) can play a significant role.

In the Southern Hemisphere, southeasterly trades dominate the entire tropical ocean equatorward of the subtropical ridge centered at 25S. The SPCZ still exists, although it is weak and rarely associated with tropical cyclogenesis.

1.2.1.4 November

As expected, the transition season exhibits a general southward movement of the major synoptic features (Fig. 2.5). The eastern North Pacific ITCZ is once again located at or south of 10N and the active portion is confined east of the col near 95W. As in May, the ITCZ/ near equatorial trough exists on both sides of the equator in the western Pacific. The SPCZ again becomes more pronounced and linked to the Australian ITCZ.

1.2.2 Indian Ocean Surface Circulation

1.2.2.1 January

The ITCZ/monsoon trough of the South Indian Ocean is very pronounced near 10S (Fig. 2.6). In the eastern portion it extends southeastward to the heat lows of the Australian continent. The North Indian Ocean is dominated by the northeast monsoonal flow which crosses the equator and converges into the ITCZ. One significant feature is the flow pattern

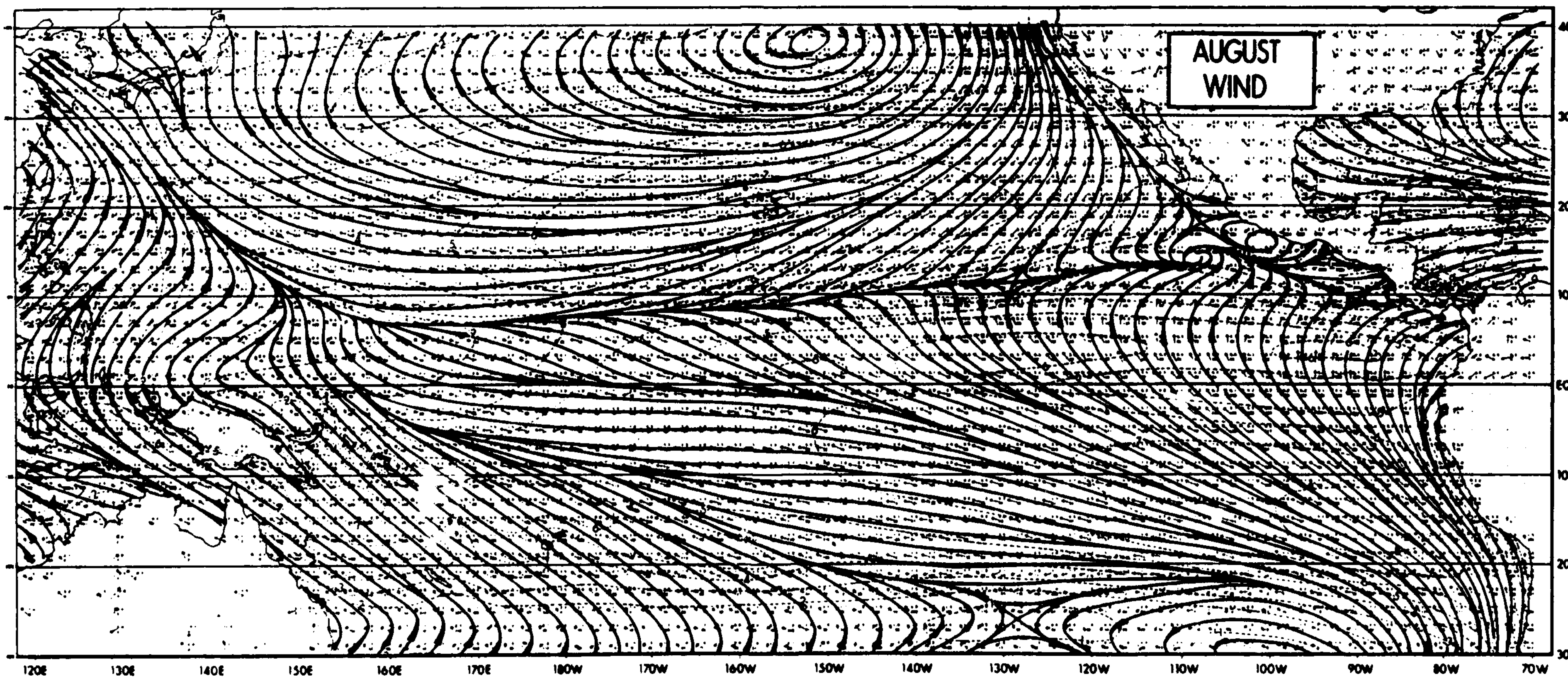

Fig. 2.4. Mean surface level streamline analyses over the Pacific for August (Sadler, 1975).

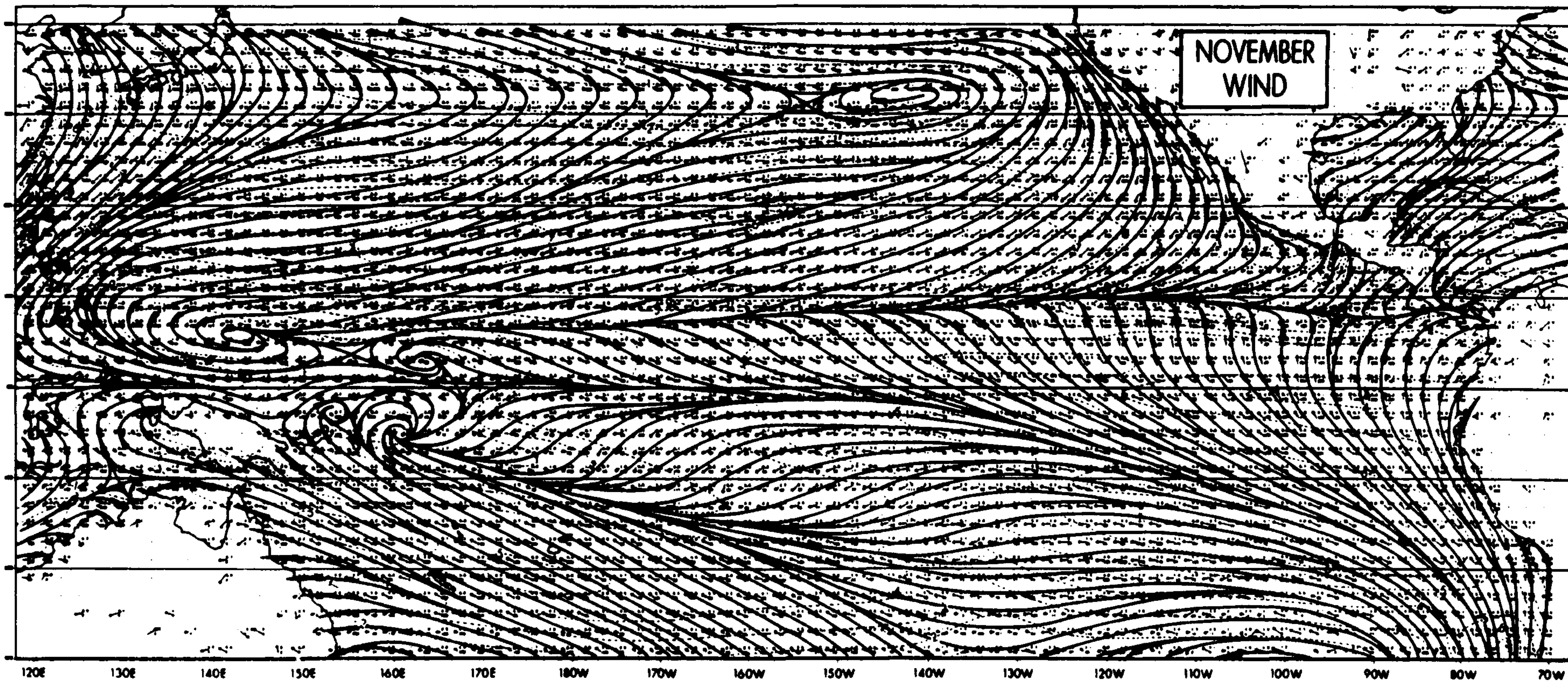

Fig. 2.5. Mean surface level streamline analyses over the Pacific for November (Sadler, 1975).

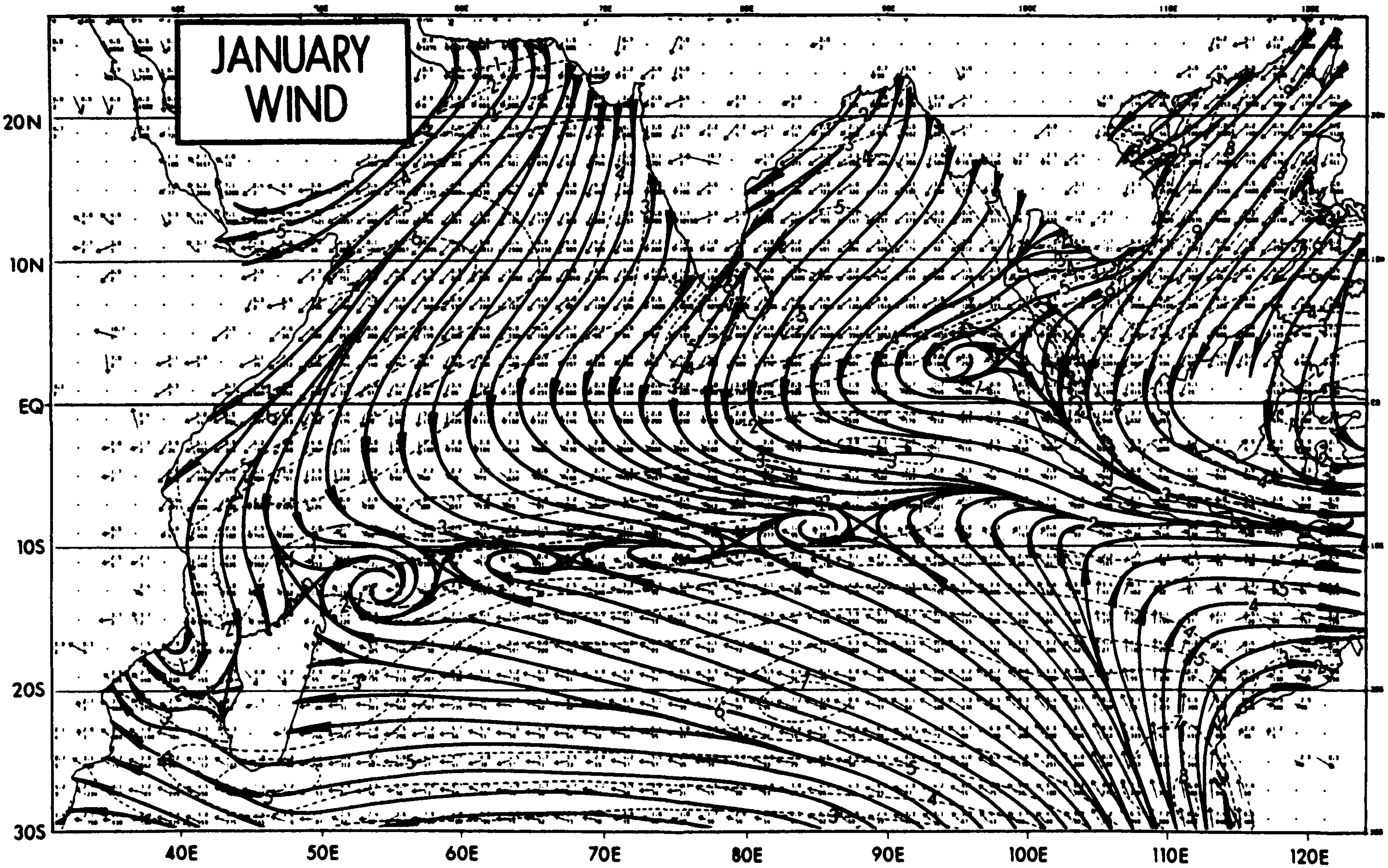

Fig. 2.6. Mean surface level streamline analyses over the Indian Ocean for January (Sadler, 1975).

in the Mozambique Channel. The southeasterly trades in this region are met by cross-equatorial northerlies creating a convergence zone in the channel. Tropical cyclones often form within the channel during this time of the year (See NTAG Vol 7, Fett et al., 1990, Section 3C, case 3, for an example).

1.2.2.2 April

During the April transition season the ITCZ/near equatorial trough exists on both sides of the equator near 5 (Fig. 2.7). Twin cyclones are not uncommon during this season in the southern Bay of Bengal. Cross equatorial flow from the north essentially ceases. A complete reversal of the January pattern takes place in May (not shown) as the southwesterly monsoon begins.

1.2.2.3 July

During July, the southwest monsoon is the dominant pattern (Fig. 2.8). Unlike the January chart, however, there is no ITCZ/trough in the North Indian Ocean. In reality, the trough has migrated northward and is located over the Indian subcontinent in the form of heat lows. The sharp turning of the winds near the equator is not a trough but rather a buffer zone. Typically it is not convectively active. In the South China Sea region, the winter northeasterly monsoon has been replaced by monsoonal southwesterlies.

1.2.2.4 October

By October, the southwest monsoon has nearly ceased (Fig. 2.9). The beginnings of the South Indian Ocean ITCZ are evident near 5S as the buffer zone has moved south from the July position. The heat lows over India have also moved south to become a trough over southern India and the open water. A month later (not shown), the two troughs are well defined at 5° north and south of the equator as in April. The northeast monsoon is once again present over the South China Sea.

1.2.3 Atlantic Ocean Surface Circulation

1.2.3.1 January

As in the eastern Pacific, cross-equatorial flow in the Atlantic is south to north throughout the year (Fig. 2.10). Southeasterly trade winds dominate the Southern Hemisphere tropics all year. A weak convergence zone exists near the equator in the Northern Hemisphere with northeasterly trades over much of the tropical North Atlantic. This convergence zone is largely inactive during the winter months.

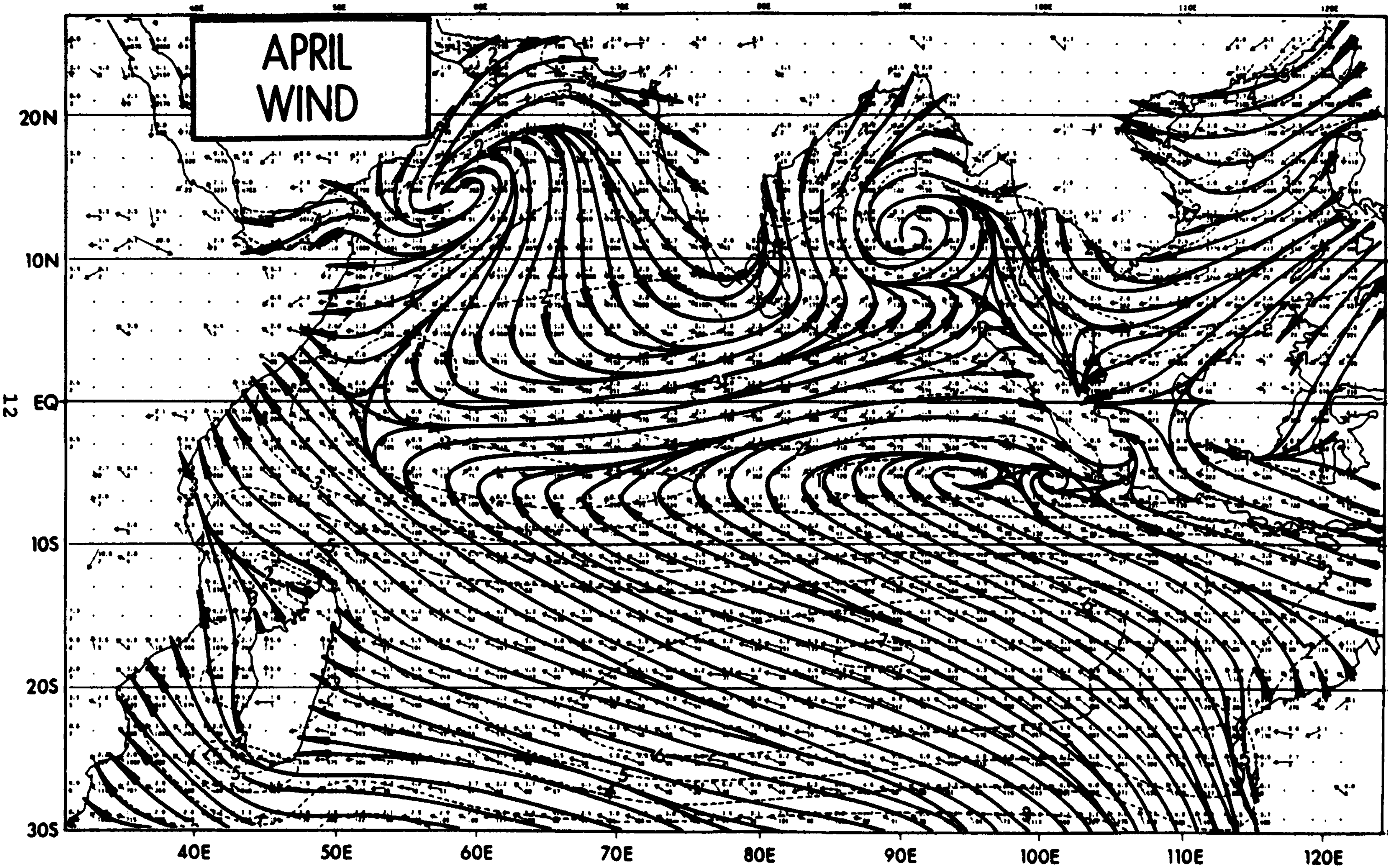

Fig. 2.7. Mean surface level streamline analyses over the Indian Ocean for April (Sadler, 1975).

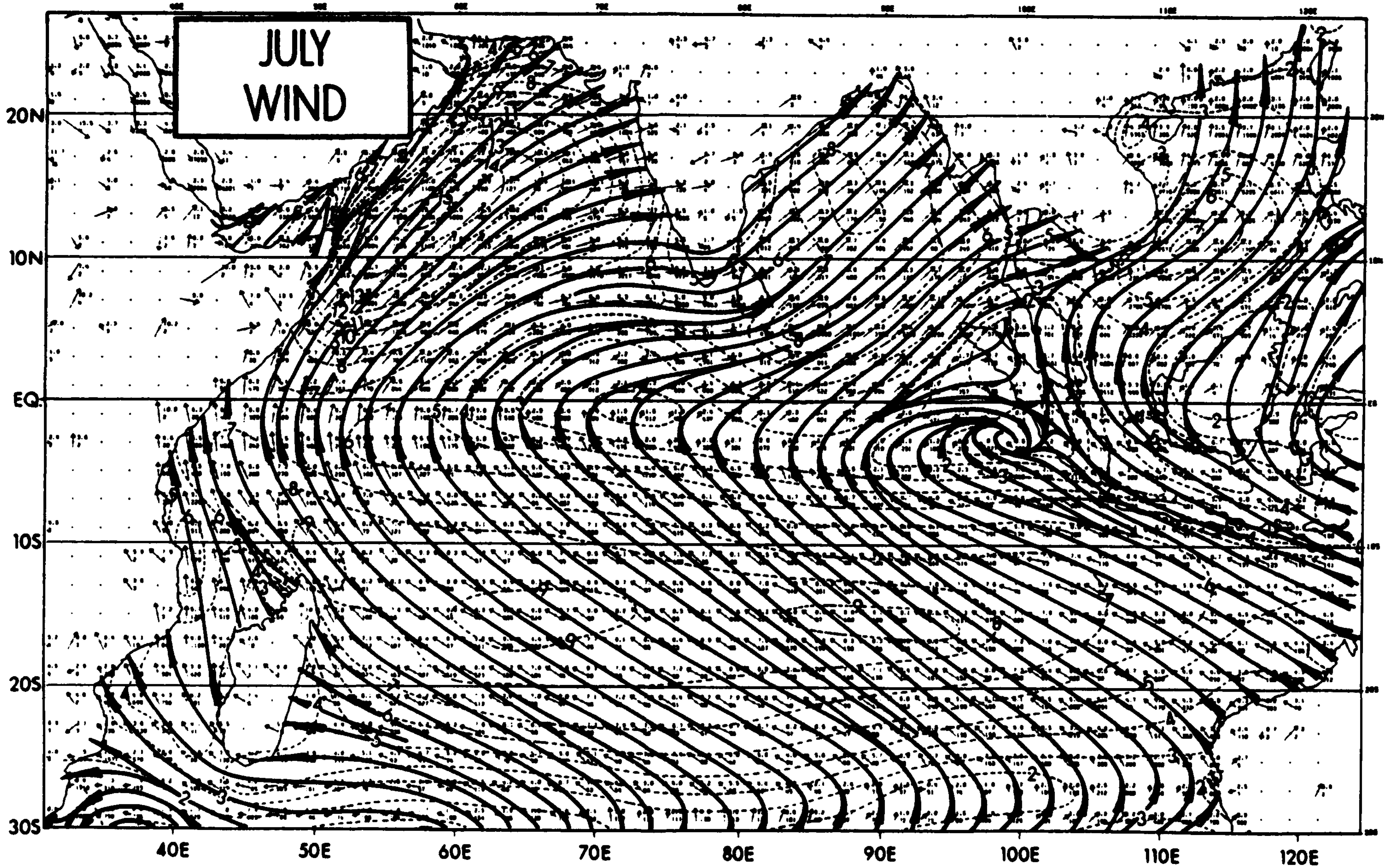

Fig. 2.8. Mean surface level streamline analyses over the Indian Ocean for July (Sadler, 1975).

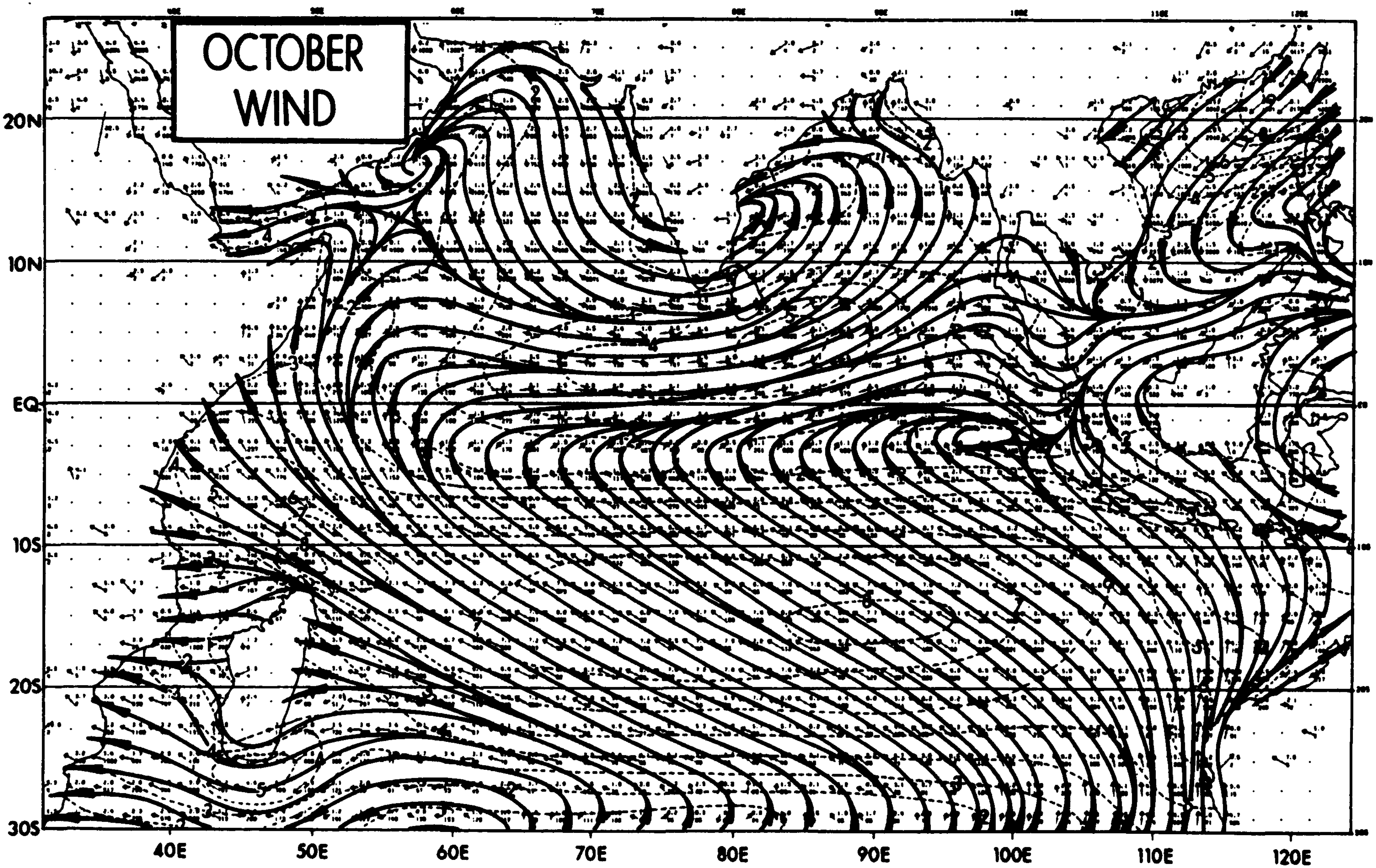

Fig. 2.9. Mean surface level streamline analyses over the Indian Ocean for October (Sadler, 1975).

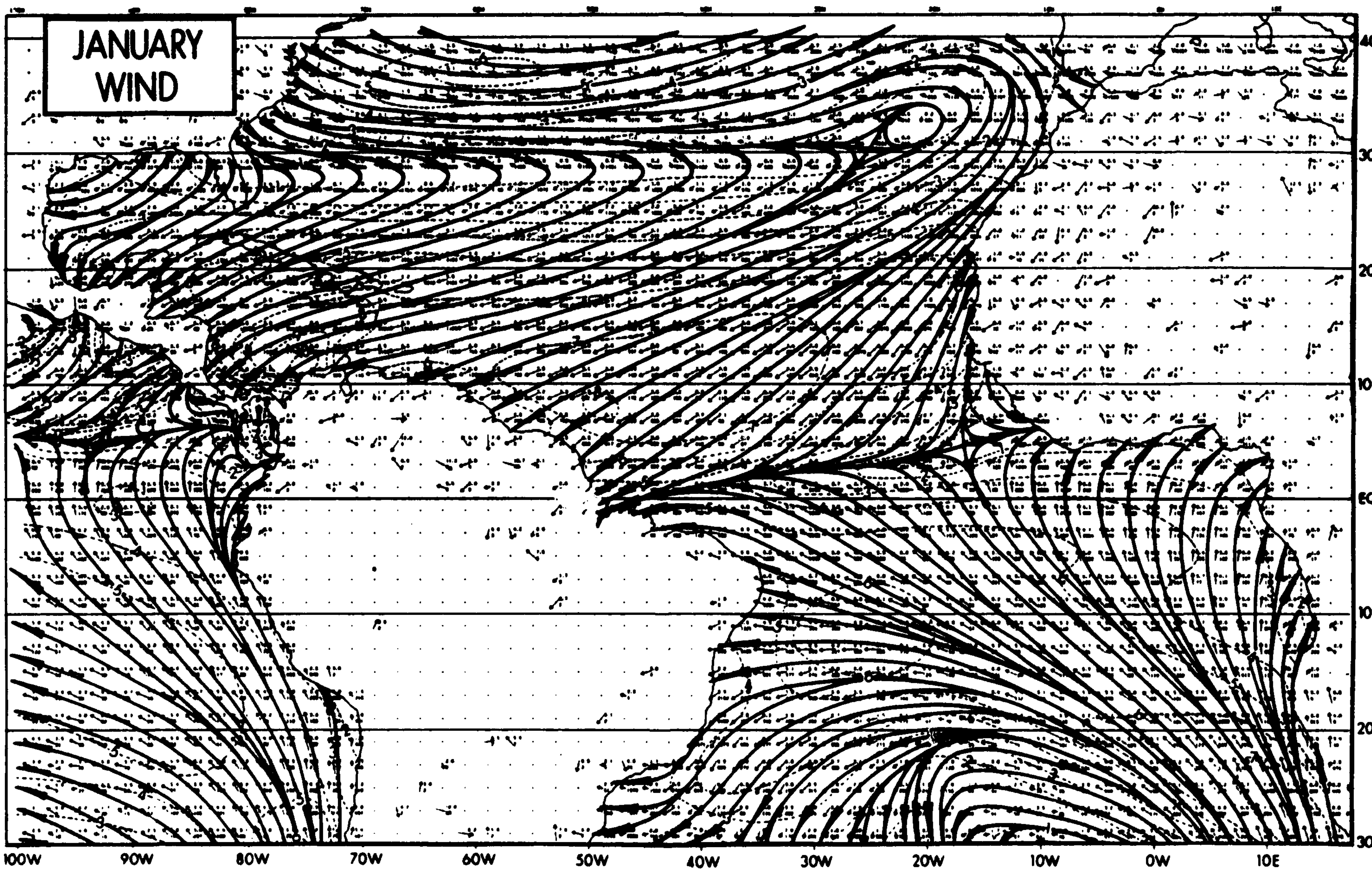

Fig. 2.10. Mean surface level streamline analyses over the Atlantic for January (Sadler, 1975).

1.2.3.2 July

The flow pattern is largely unchanged from February (Fig. 2.11). The most significant feature is the movement of the ITCZ northward to near 10N. As in the eastern North Pacific, the ITCZ displays a convergent and confluent portion east and west of the col at 40W respectively. In the Gulf of Mexico, winter northeasterlies have been replaced by summer southeasterlies.

1.2.4 Tropical Upper Air Circulation

1.2.4.1 January

Strong westerlies dominate the Northern Hemisphere winter. Three speed maxima are observed; one over Japan (Fig. 2.12), one over North Africa (Fig. 2.13), and one over the east coast of North America (Fig. 2.13). These maxima are located where the polar and subtropical jet streams typically merge. Semi-permanent anticyclones cover much of the Southern Hemispheric tropics. Over the western Pacific and Indian Ocean, the outflow from these anticyclones crosses the equator (Fig. 2.12). This outflow serves as the upper-level branch of the a compensating circulation for the low-level flow from north to south. The Southern Hemispheric TUTT is also apparent, extending from 30S, 110W to 0°, 180°.

1.2.4.2 July

Two prominent upper-level anticyclones exist in the Northern Hemisphere; one over North America (Fig. 2.14) and one over the Tibetan Plateau (Fig 2.15). The latter anti-cyclone is massive in size and is a product of the Indian summer monsoon. TUTT's exist in both the Pacific and Atlantic. The Pacific TUTT starts in the Gulf of Alaska and stretches southwest and then west into the Philippine Sea (Fig. 2.15). In the Atlantic, the TUTT extends from west of the United Kingdom into the Gulf of Mexico (Fig. 2.14). At times this TUTT stretches across Mexico and into the Pacific Ocean.

The cross-equatorial flow is entirely from the Northern to Southern Hemisphere. This upper-level flow pattern is consistent with the positions of the surface monsoon troughs/ ITCZ, which all lie north of the equator at this time of year. Strong wintertime westerlies dominate the Southern Hemisphere south of 20S with three maxima: one over Australia (Fig. 2.15), one west of South America (Fig 2.14), and one over Africa (Fig. 2.14).

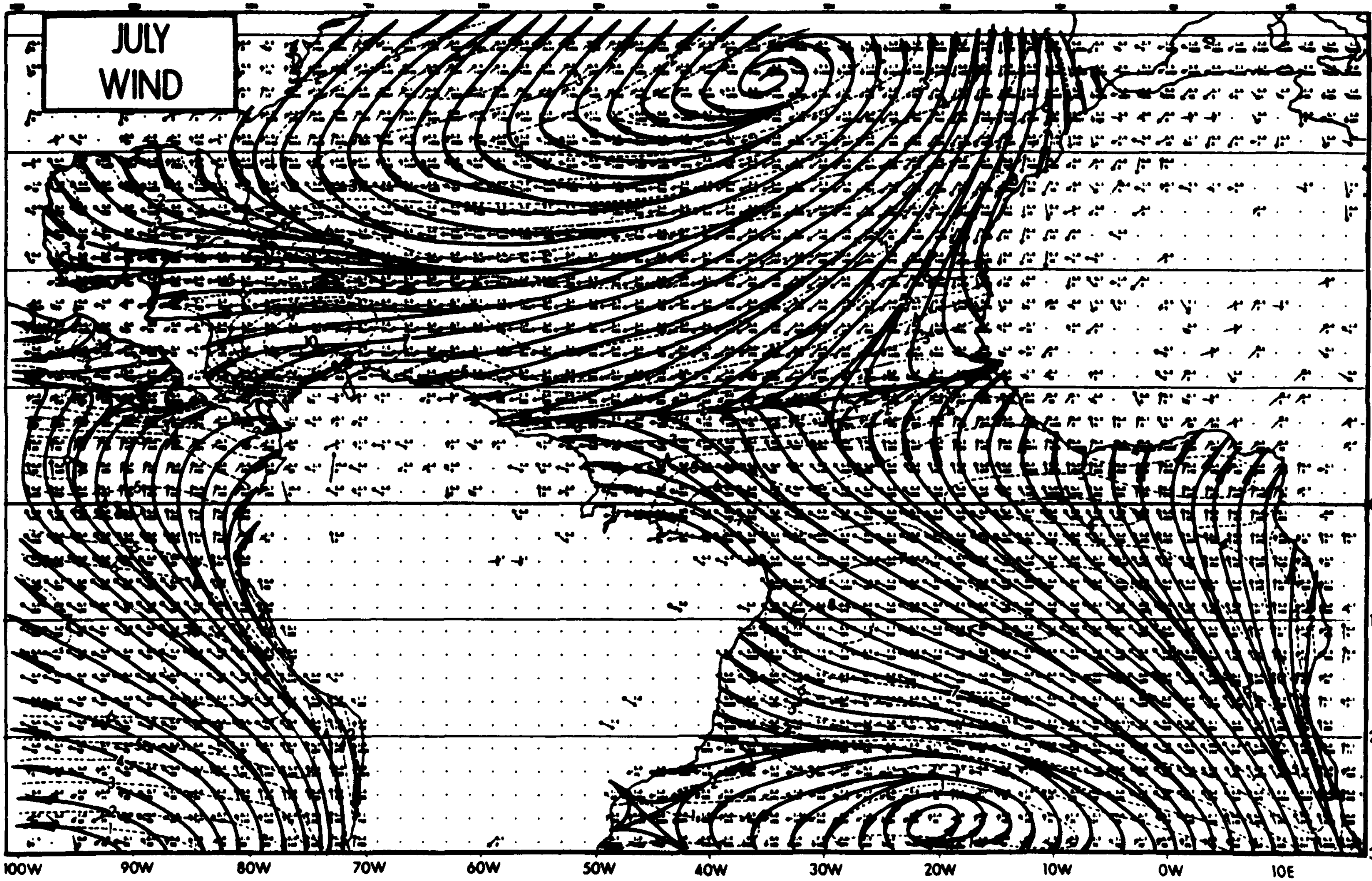

Fig. 2.11. Mean surface level streamline analyses over the Atlantic for July (Sadler, 1975).

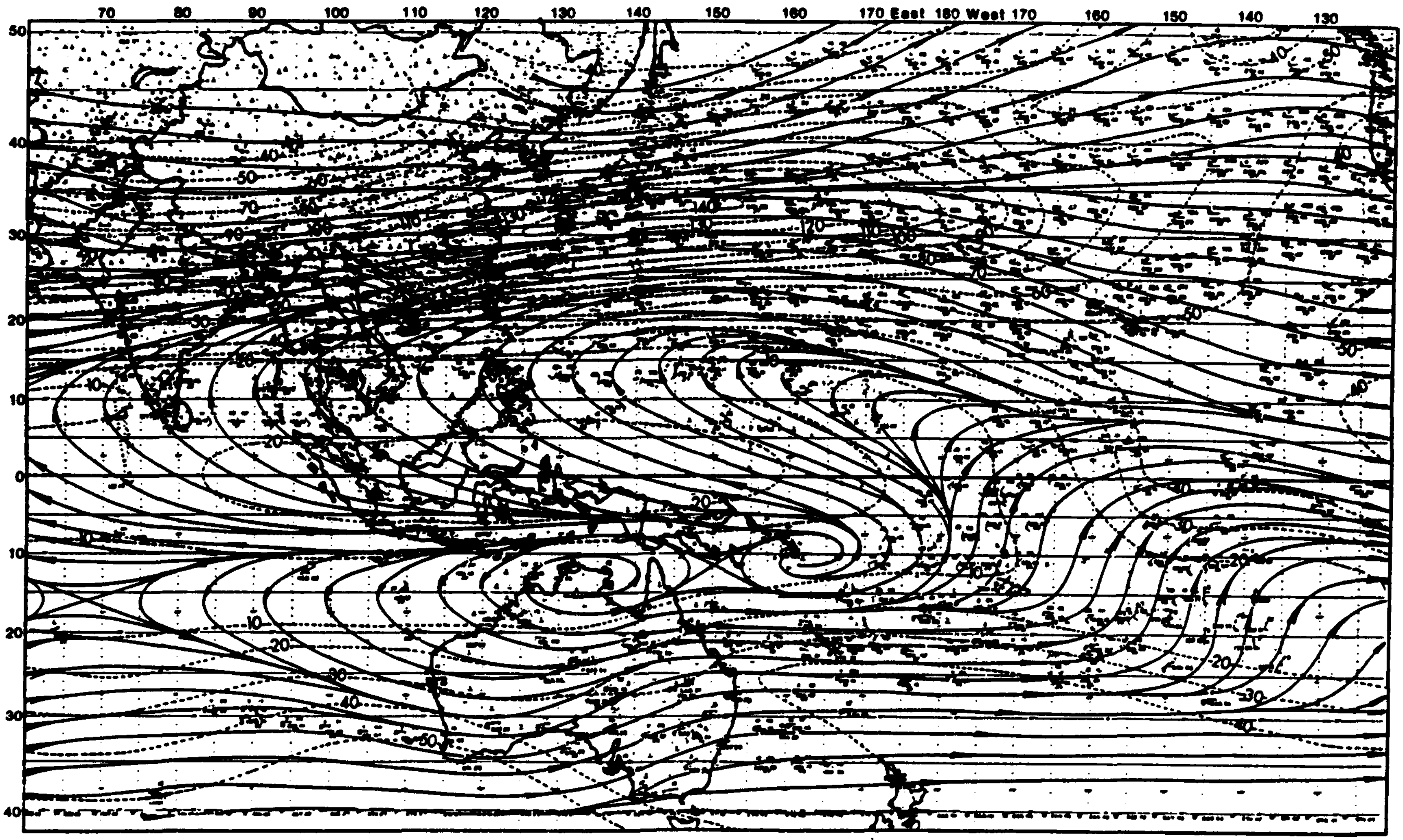

Fig. 2.12. Mean 200 mb level streamline analyses over the Pacific for January (Sadler, 1975).

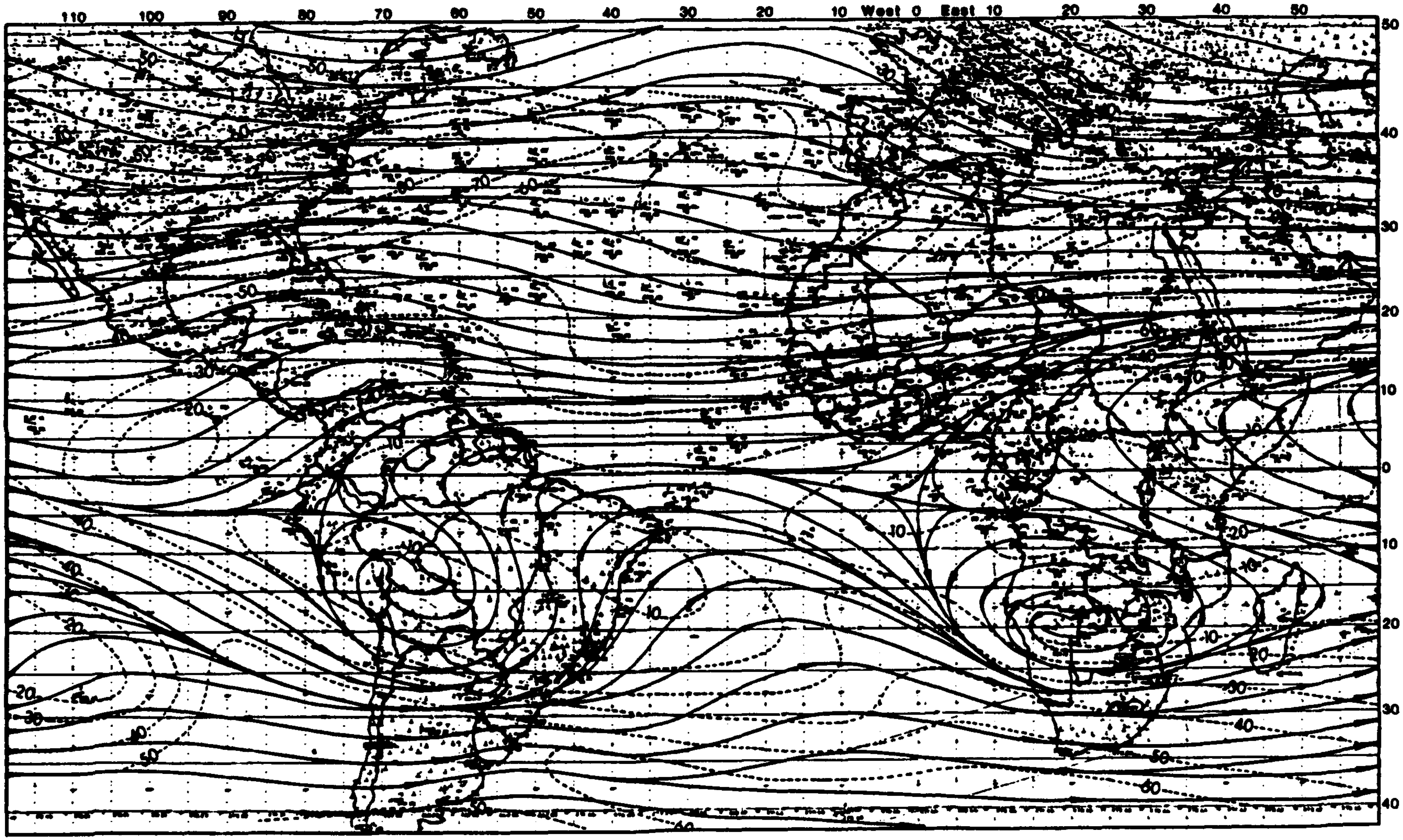

Fig. 2.13. Mean 200 mb level streamline analyses over the Atlantic for January (Sadler, 1975).

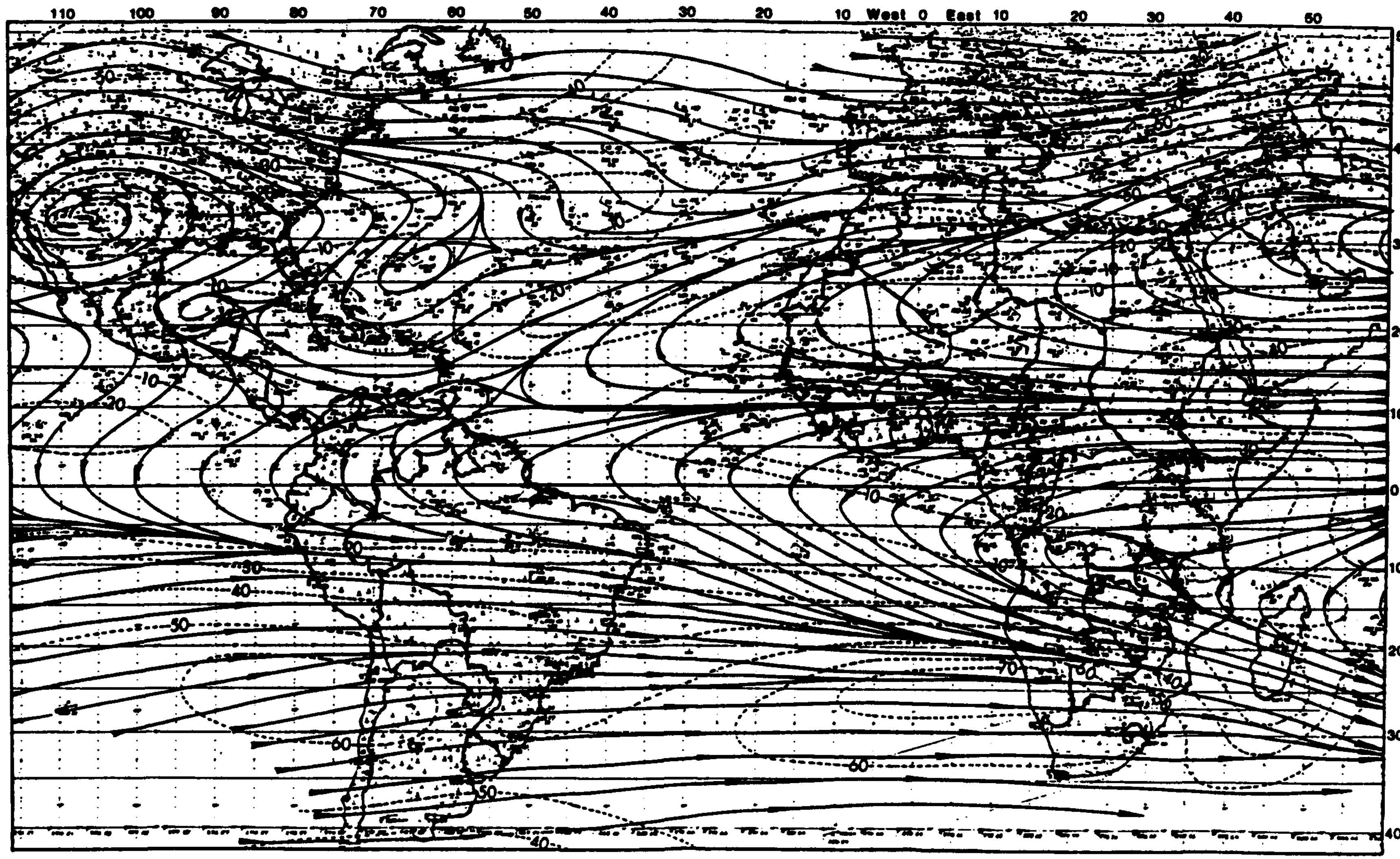

Fig. 2.14. Mean 200 mb level streamline analyses over the Pacific for July (Sadler, 1975).

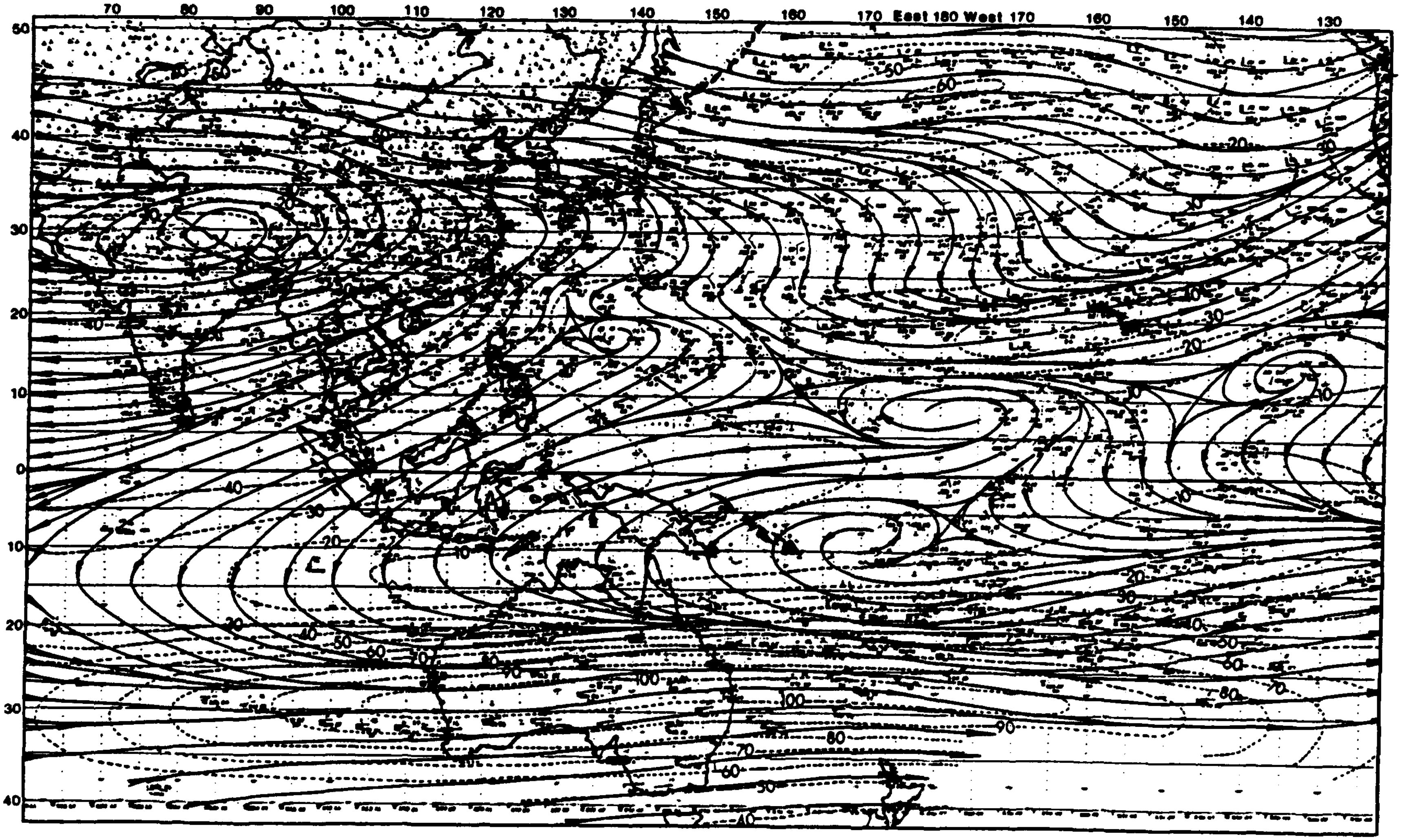

Fig. 2.15. Mean 200 mb level streamline analyses over the Atlantic for July (Sadler, 1975).

1.2.5 Vertical Motion in the Tropics

In the tropics three dimensional circulations also exist. Figures 2.16 shows seasonal mean vertical motion at the 500 mb level (Newell et al., 1974). In general, there is large-scale ascending motion ($W < O$) in near-equatorial latitudes and descending motion ($W > O$) in the subtropical latitudes. Note, however, that zonal variations of large-scale vertical motion can be greater than meridional variations. Areas of large vertical motion such as the Amazon and Indonesia indicate deep convection and heavy rain (Panofsky, 1951).

1.2.6 Mean Zonal Winds in the Tropics

Figure 2.17 shows the latitude-altitude cross section of mean zonal wind speed for the four seasons. The positive and negative values indicate westerly and easterly wind components, respectively. Some prominent features are:

1.2.ᴗ.1 Subtropical Jetstreams

These tropospheric (below 100 mb level) westerly wind maxima, located near the 200 mb level, have seasonal variations in terms of magnitude and location.

1.2.6.2 Easterly Wind Regime

The mean easterlies in the topics have two maxima in the vertical, one is located near the earth's surface, the other in the upper stratosphere.

1.2.6.3 Vertical Shear

In the troposphere mean zonal wind speeds generally increase with altitude; thus, vertical wind shear is positive. This vertical shear is strongest in the middle and higher latitudes. Weaker vertical shear is found in the tropics.

1.2.7 Significant Seasonal Characteristics of Mean Zonal Wind in the Troposphere

1.2.7.1 December - February

a. The northern hemisphere mid-latitude jet reaches its seasonal maximum intensity with a mean speed 70 knots or 35 m/s around 30N.

b. The surface easterly maximum (10 knots or 5 m/s) at 10N is about two-and half times as strong as the one near 24S.

c In the tropics, there are mean westerlies in the upper troposphere except in the equatorial zone from 3N to 10S.

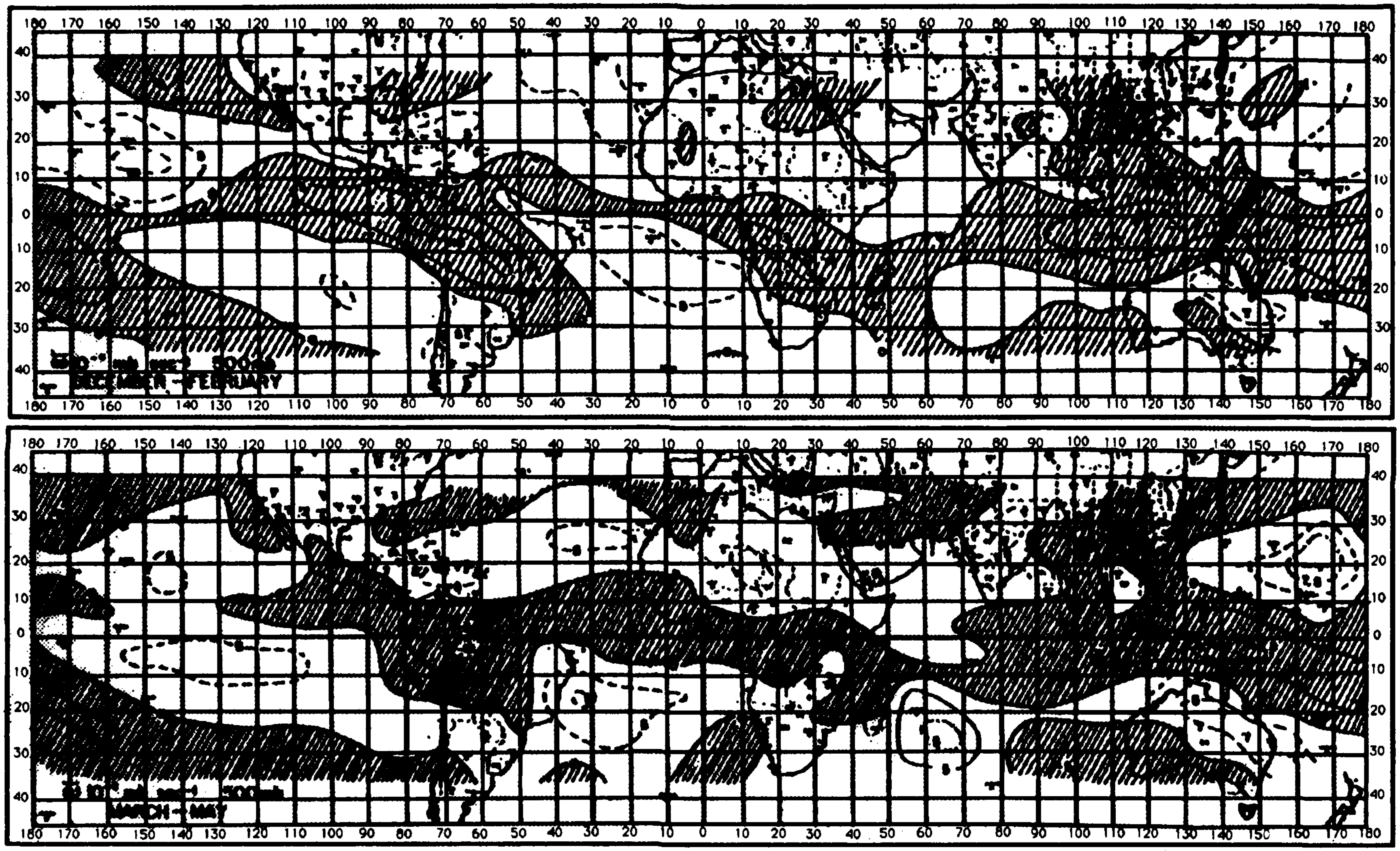

Fig. 2-16. General circulation of the tropical atmosphere and interactions with midlatitudes (Newell, 1974).

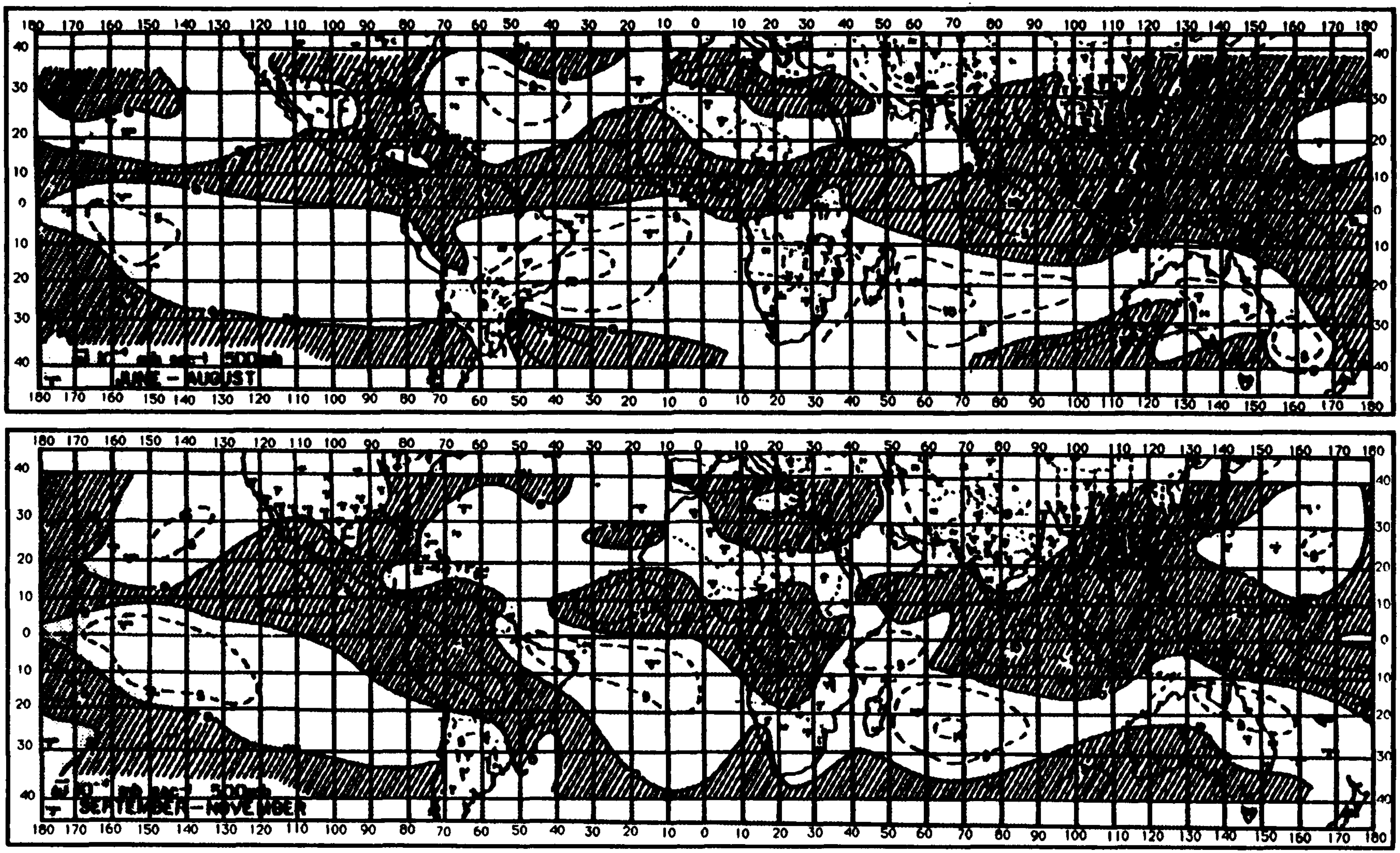

Fig. 2.16, Continued.

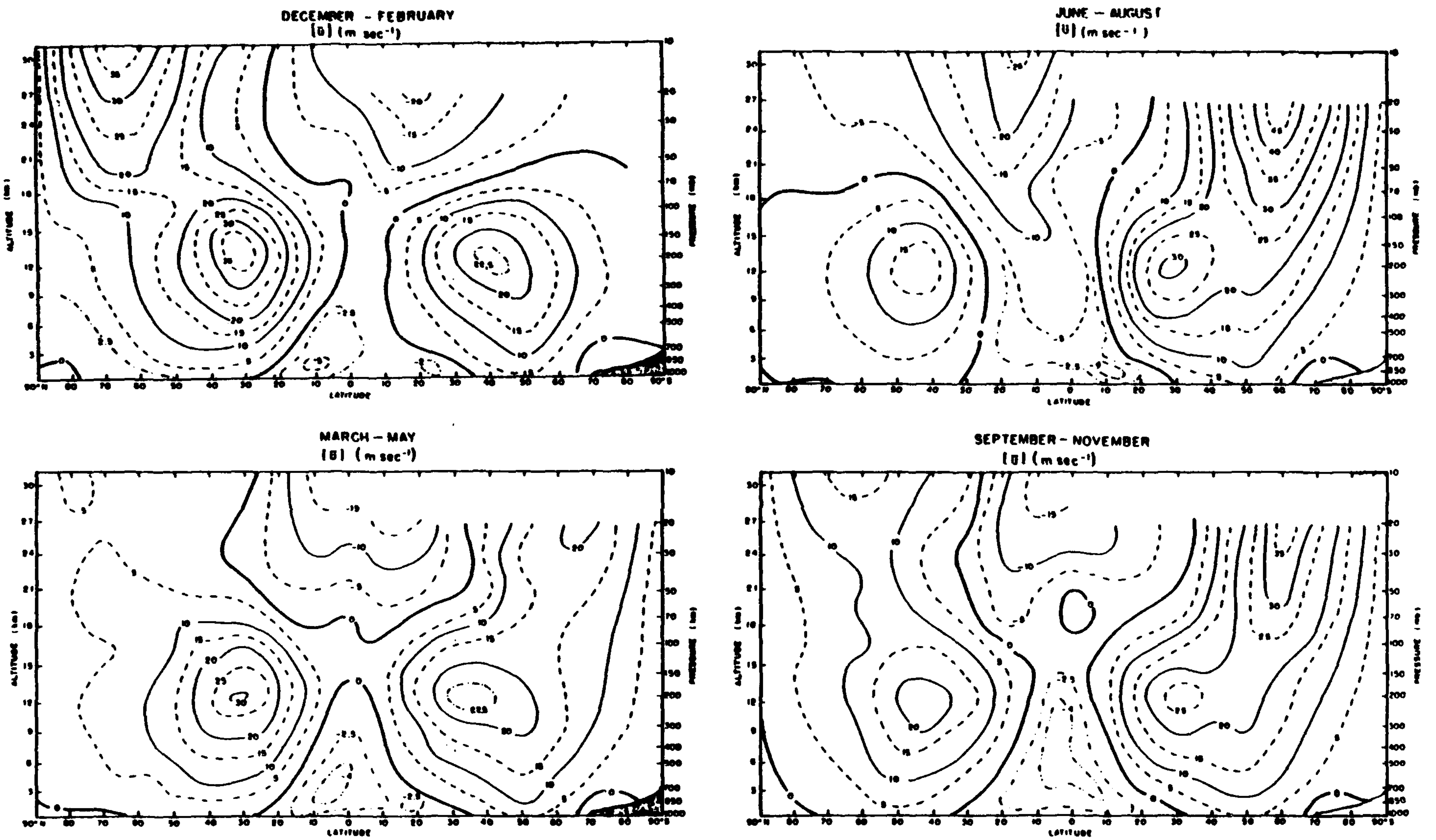

Fig. 2.17. The latitude/altitude cross sections of mean zonal wind speed for the four seasons. Positive and negative values indicate westerly and easterly wind components, respectfully (Newell, 1074).

1.2.7.2 March - May

a. The tropical easterlies are nearly symmetrical with respect to the equator.

b. The intensity of the northern hemisphere jet decreases from the previous period.

1.2.7.3 June - August

a. The Northern Hemisphere subtropical jet reaches its seasonal minimum intensity around 45N.

b. The southern hemisphere subtropical jet reaches its seasonal maximum intensity.

c. The tropical easterlies becomes more intense than in the last season, and occupy a wider and deeper zone. The mean speeds increase with height. The surface maximum (6 kt or 3 m/s) is at 15S.

1.2.7.4 September - November

a. The northern hemisphere jet has increased in intensity.

b. The tropical easterlies have decreased in the middle and upper troposphere from the previous season.

1.2.8 Horizontal Pressure Gradients in the Tropics

Large-scale horizontal pressure gradients in the tropics are usually flatter than in higher latitudes. This can be attributed to the low values of Coriolis in the tropics and the effect it has on large scale horizontal pressure gradients. In the tropics, it is the wind field that contains the most information. Therefore, streamline and isotach analyses become key charts in tropical weather analysis and forecasting.

Figure 2.18 shows the latitude-altitude cross section of mean mass flux for the four seasons. Mass flux is the rate of transfer of mass through an area and is positive when the flux is upward or poleward. A pattern of alternating positive and negative mass flux is seen in the figure and defines some major features:

a. There are three cellular patterns evident in each hemisphere.

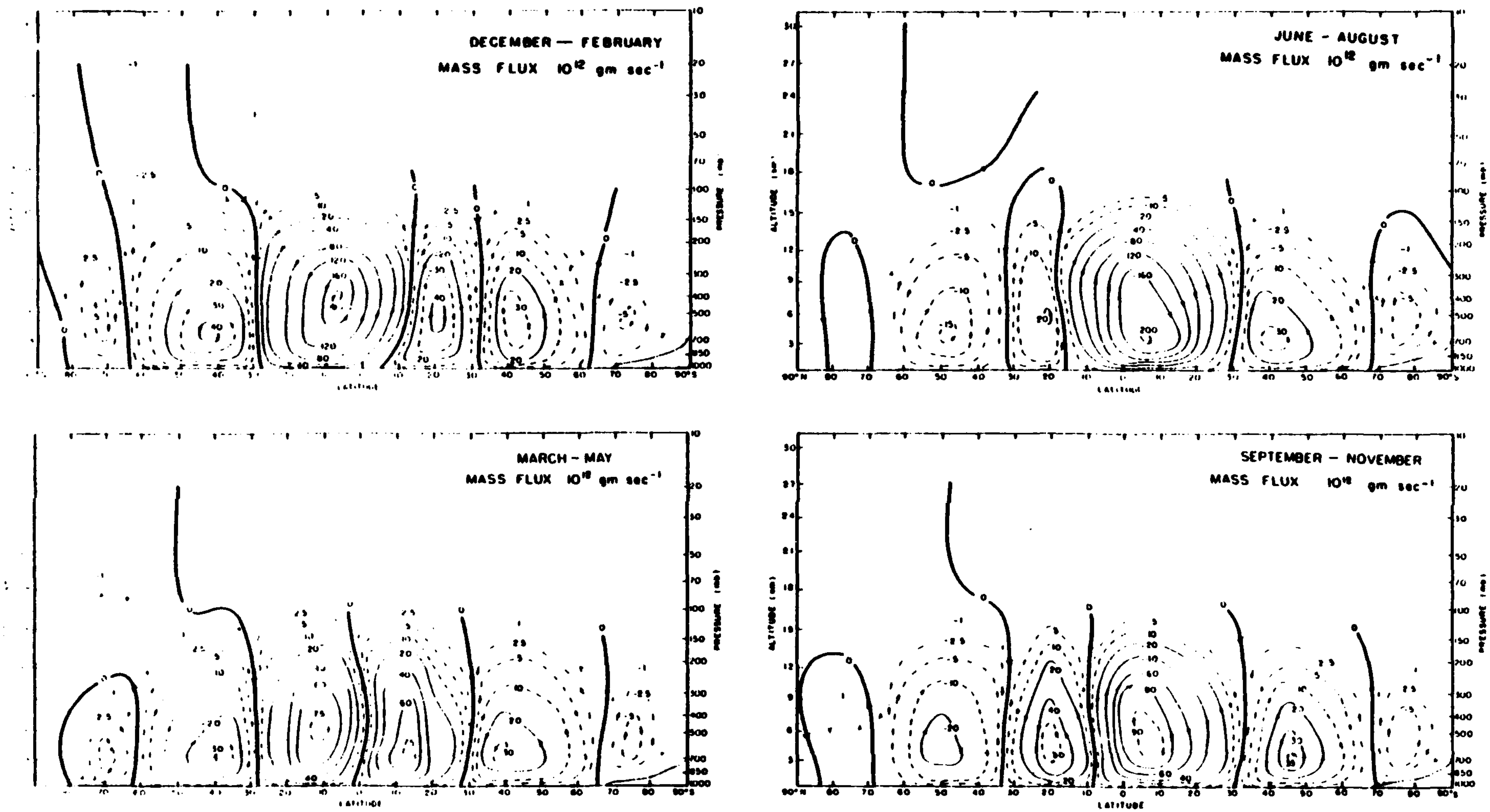

Fig. 2.18. The latitude/altitude cross sections of mean mass flux for four seasons (Newell, 1974).

b. The cells with rising motion in near-equatorial latitudes and sinking motion in the subtropics are known as the Hadley circulation (Huschke, 1957). The circulation is strongest during the winter seasons with the upward motions extending across the equator into the opposite hemisphere.

c. The cells with rising motion around 60° latitude and sinking motion in the subtropics are known as the Ferrel circulation (Fairbridge, 1967). The circulation is strongest during the winter seasons.

d. The Hadley and Ferrel circulations carry heat, moisture and momentum with them, and are responsible for the pole ward transport of energy.

1.2.9 Horizontal Temperature Gradients in the Tropics

Figure 2.19 shows the latitude-altitude cross section of mean temperature for four seasons. Some major features are:

a. The warmest temperatures in the lower troposphere and the coldest temperatures aloft are found in the tropics. This indicates that the air is generally less stable and hence amenable to convective activity.

b. At the 200 mb level, a warmer zone is located near 50° latitude. The sign of horizontal temperature gradient changes across this warmer zone, which indicates a temperature maximum located in the area.

c. In the troposphere, the meridional temperature gradients in the middle and higher latitudes are steeper than in the tropics.

d. At the earth's surface, the zone of warmest temperature is located in the summer hemisphere. This zone is known as thermal equator. The thermal equator lags behind the solar equator by a few months because heating is still occurring in the summer hemisphere when the sun is past summer solstice.

1.2.10 Specific Humidity Distribution

Figure 2.20 and 2.21 show the latitude-height cross section of mean specific humidity, a measure of water vapor content in the atmosphere, for January and July. The tropospheric water vapor in the tropics has smaller seasonal variations than that in the middle and higher latitudes. The vertical decrease of water vapor content in the tropics is greater than that in the other latitudes. The combined effects of large vertical temperature and water vapor gradients make the tropical atmosphere conditionally unstable. This means that there

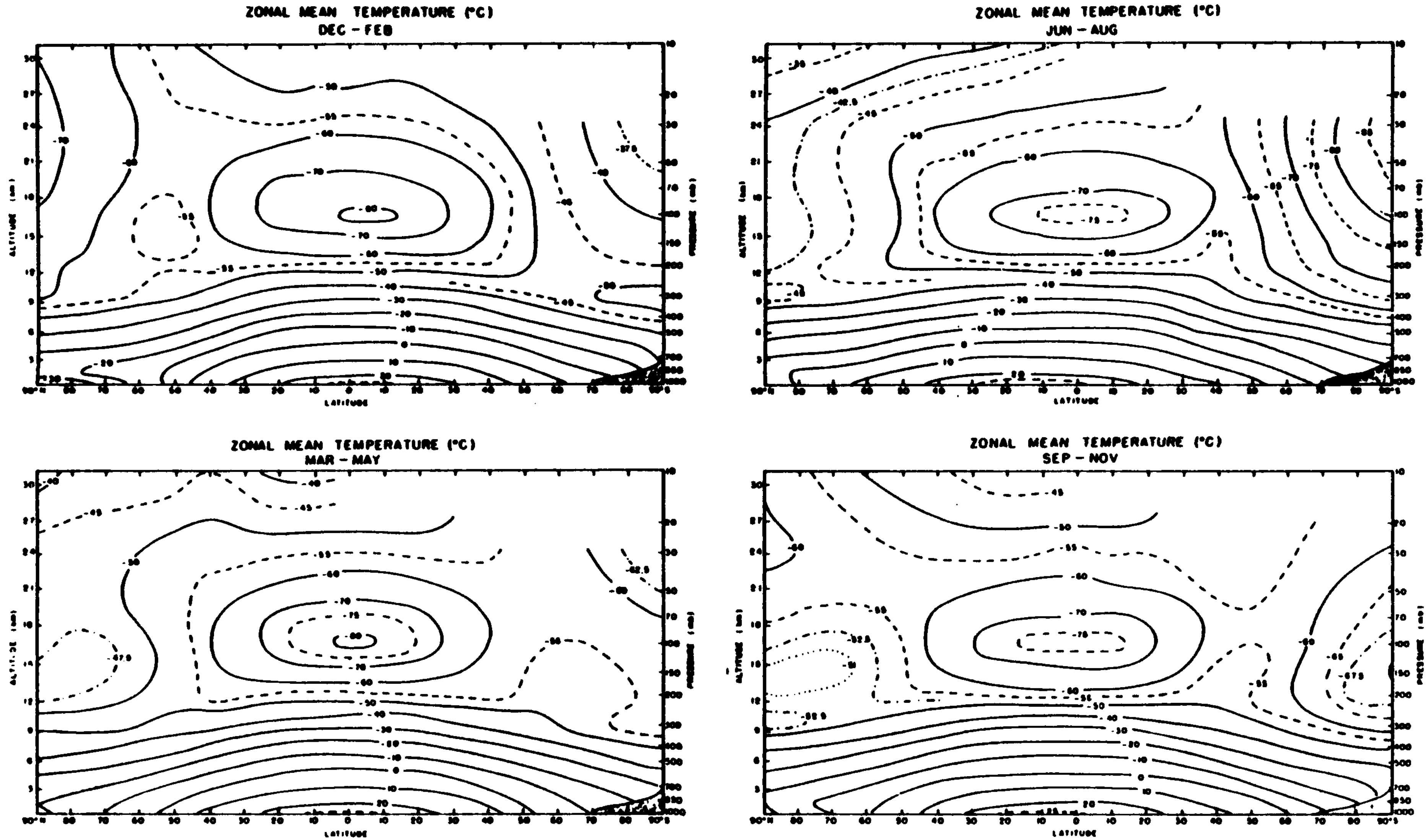

Fig. 2.19. The latitude/altitude cross sections of mean temperature for four seasons (Newell, 1974).

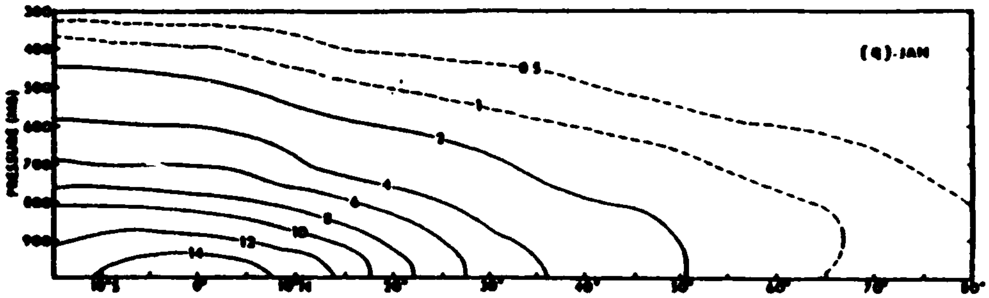

Fig. 2.20. Longitudinally averaged specific humidity for January. Units: gm/kg

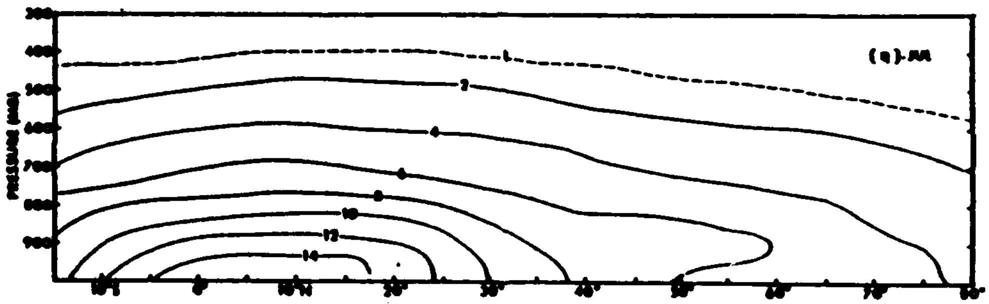

Fig. 2.21. Longitudinally averaged specific humidity for July. Units: gm/kg

is a height to which a parcel of air can be lifted such that it will continue to rise on its own because it is more buoyant than the surrounding air.

1.2.11 Radiative Heating Rates

Figure 2.22 shows the latitude-height cross section of mean radiative heating for each season. Note that most of the atmosphere undergoes radiative cooling (negative values) with the exception of a large region of radiative heating (positive values) in the tropics above 150 mb. Maximum cooling is found in the tropics near 300 mb where cooling due to the presence of water vapor more than compensates for the effects of heating due to solar absorption. Cooling is also found near the surface between the equator and 20N. Radiative heating in the tropics has less seasonal variations than that in the midlatitudes. Vertical distributions show cooling at the surface and above the 500 mb level.

1.3 Physical Factors

The observed large-scale circulations were discussed in the previous section. Physical processes of the circulations are described in this section.

1.3.1 Local Effects

1.3.1.1 Distribution of Land and Water

a. Land/water distribution - The distribution of land and water has a significant effect on the entire tropical atmosphere. Radiative heating and cooling of the land masses of Africa, Asia, Indonesia, Australia and South America cause the monsoonal flow over much of the tropics. Large scale monsoon troughs develop in the flow with westerly winds on the equatorward side and easterlies on the poleward side. There is a direct correlation between the locations of these troughs and the development regions of tropical cyclones.

b. Effect of the Tibetan Plateau - The Tibetan Plateau has been identified as an elevated source region where considerable latent heat of condensation is released into the atmosphere (Flohn, 1957; He et al., 1987). This middle tropospheric heat source is the primary development mechanism for the Asian southwest monsoon or summer monsoon. The summer monsoon is best identified with the development of an anticyclone on 200 mb streamline chart during summer. As a result of the mid-level heating an easterly jet develops equatorward of this anticyclone. This jet further contributes to the widening easterly regime in the middle troposphere and the development of the southwest monsoon.

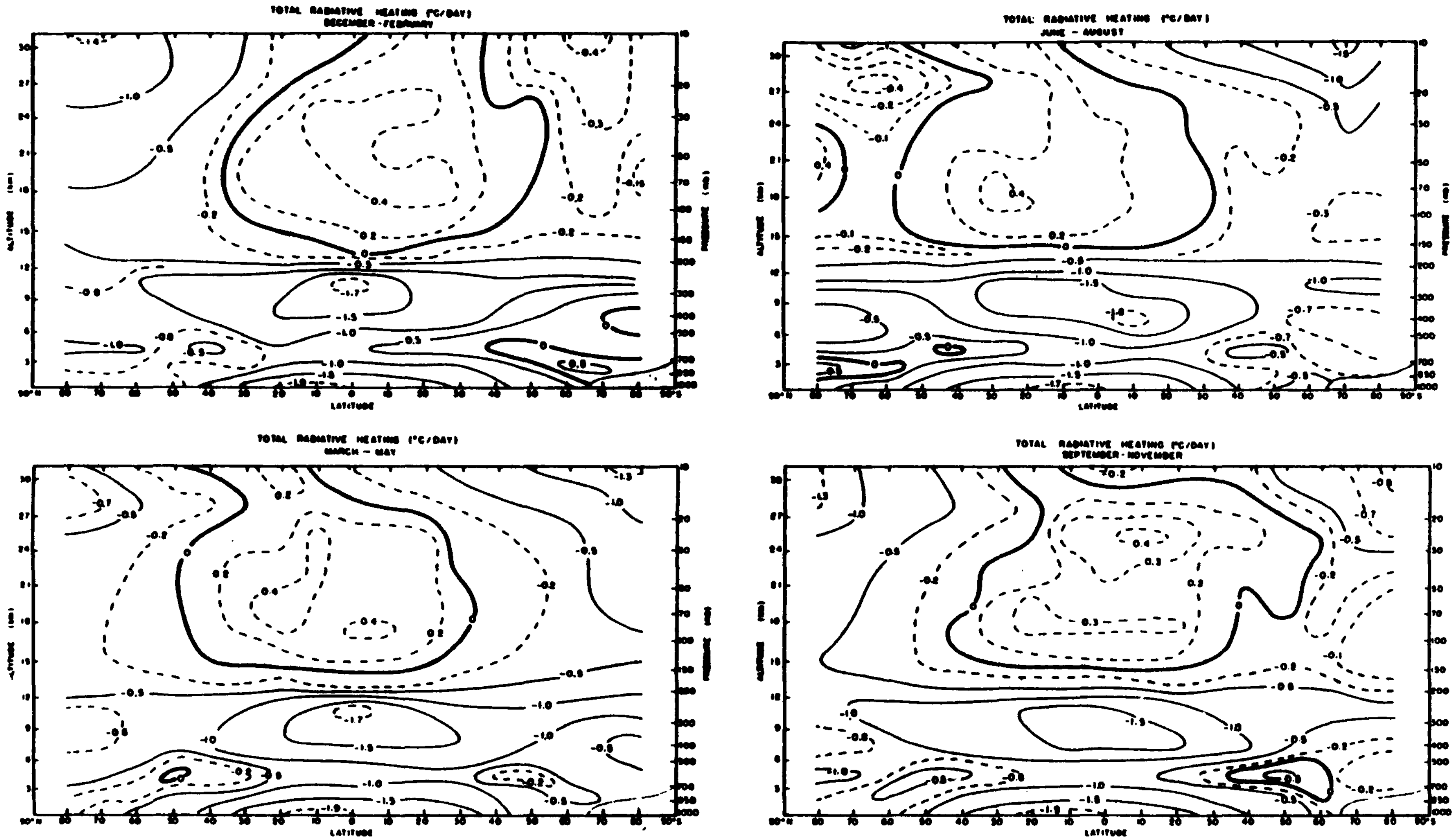

Fig. 2.22. The latitudinal distribution of the total radiative heating rate (C/day). This is the sum of all contributions from both solar and infrared radiation transfer processes (Newell, 1974).

1.3.1.2 Sea-Surface Temperature Effects

a. Dynamic contributions - Sea surface temperature and sea state are the two key factors which determine the heat, moisture and stability across the air-sea interface on a daily basis.

b. Tropical cyclone development contributions - Palmen (1948) studied the thermodynamic conditions of North Atlantic hurricanes. He concluded that all oceanic regions with an air temperature above 27.5°C are within the regions of hurricane origin. After formation, hurricanes usually move out of the warm regions into higher latitude. He also concluded that hurricanes can only form in the oceanic regions away from the equator due to the lack of Coriolis force.

c. Tropical Cyclone SST Modifications - Tropical cyclones can modify the distribution of sea-surface temperature. Leipper (1967) conducted a case study of the effects of hurricane Hilda on the sea surface temperature in the Gulf of Mexico. He found that the warm ocean surface layers were transported outward from the hurricane center, while cold water upwelling along the storm track was about 60 meters deep. The upwelling caused sea surface temperatures to decrease more than 5°C over an area of 70 to 200 miles wide.

d. SST/Rainfall Amounts/Walker Cell Theory - There is a direct relationship between sea-surface temperature and rainfall amounts in the tropics. Bjerknes (1966, 1969) hypothesized that the occurrence of warmer sea surface temperatures in the east and central equatorial Pacific are associated with weakening trade winds and decreased equatorial upwelling. His model shows that the Walker circulation has rising motion concentrated near the dateline and subsidence in the eastern Pacific. Another model (Fig. 2.23) shows a similar circulation with upward vertical motion over Southeast Asia and subsidence over the Sahara.

1.3.1.3 Tropical Weather Indices

a. The southern oscillation Index - The Southern Oscillation describes conditions associated with low temperatures in the Pacific and Indian Oceans and rainfall amounts that vary inversely with pressure. The Southern Oscillation Index is defined as a function of the surface pressures at Santiago, Honolulu, Manila and Darwin, and the rainfall in India, Chile, and the Nile Flood. Walker and Bliss (1932) used the index to prepare

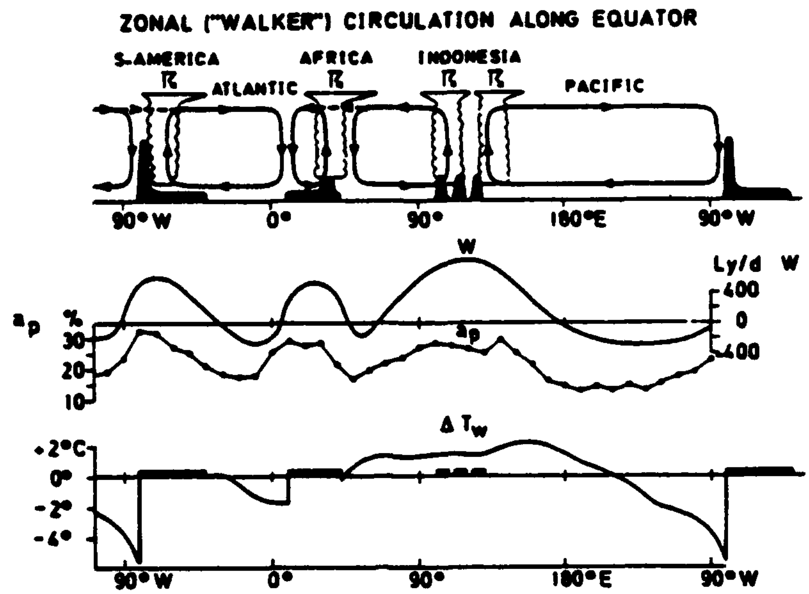

Fig. 2.23. Zonal (Walker) circulation along the equator. Symbols are w, heat budget of an atmospheric column; ap, planetary albedo; Δ tw, departure of sea surface temperature from the latitude of mean (Flohn, 1071).

seasonal forecasts of rainfall and pressure in India. A simplified version of the southern oscillation index can be determined from the difference between the mean monthly sea level pressure at Easter Island and Darwin, Australia.

Troup (1965) found that when there is a high southern oscillation index (i.e., greater than +1 standard deviation) during the period from June to August, the Indian Ocean southeast trades weaken and the Arabian Sea branch of low-level westerlies strengthen. Other research indicates there may be some correlation between the Southern Oscillation Index and the El Nino.

b. The El Nino - The El Nino, (meaning the Child in Spanish), is a unique example of large-scale sea surface temperature effects on the weather (Bjerknes, 1959, 1961).

(1) During normal years, the Peru Current moves northward from the very cold waters of the Antarctic Circle. The Coriolis effects on the current combined with offshore winds tend to deflect the Peru Current away from the shore, resulting in upwelling. This regular phenomenon occurs from late December till March.

(2) During abnormal years, a warm equatorial countercurrent becomes very strong along the coast and causes displacement of the cold Peru Current away from the coast. Such changes in ocean currents can cause catastrophic changes in the plankton, fish, bird and human population.

(3) The opposite mode of El Nino with an occasional cold tropical ocean current is called La Nina.

(4) Rasmusson and Carpenter (1982) pointed out that the term El Nino has been used to encompass the large-scale features of warming episodes associated with upwelling along the equator and the South American coast. The study determined that 67 out of 71 cases since 1803 have a repeating interval with a range of between 1 and 5 years.

1.3.1.4 ENSO

The El Nino and southern oscillation have been treated as one system; this system is commonly called ENSO. Rasmusson and Carpenter (op.cit.) developed a composite picture of the five phases of ENSO; they are pre-onset, onset, peak, transition, and mature phases.

a. Pre-onset Phase - This occurs during August-October of the year preceding El Nino. The sea surface temperatures in the central and eastern equatorial Pacific (about 150E-080W, 15S-15N) are about 0.2° to 0.4°C colder than normal. The surface winds are westerly in the vicinity of the equator, west of the dateline.

b. Onset Phase - This occurs during November-January prior to the maximum sea surface temperature on the Ecuador-Peru coast. The sea surface temperature around the dateline (170E-170W) becomes 0.2° to 0.4°C warmer than normal. In the same area, the easterly surface winds are replaced by a westerlies. Most of the eastern equatorial Pacific (east of 170W) is slightly colder (about 0.2°C) than in the pre-onset phase.

c. Peak Phase - This occurs during March-May of El Nino year. Almost the entire tropical Pacific Ocean is anomalously warm, with the maximum (greater than 1°C) located off the coast of Peru and Chile. The ITCZ in the eastern North Pacific is displaced south of its normal position, but still north of the equator. The westerlies of the western, central and eastern Pacific have increased from those of the onset phase. The trades over the entire Pacific become very weak.

d. Transition Phase - This occurs during August-October of the El Nino year. The warm sea surface temperatures prevail in the entire equatorial Pacific from 160 E to the South American coast, with a zone of maximum temperature along the equator. The ITCZ is located north of the maximum sea-surface temperature zone at about 5-10 degrees latitude. In the central Pacific (160E-130W), the maximum sea-surface temperature zone and the low-level convergence zone are colocated. The westerlies in the western equatorial Pacific (130E-150W) are well established.

e. Mature Phase - This occurs during December-February following the El Nino. The teleconnections between the tropics and the Northern Hemisphere subtropics are strongest during this phase. The temperatures around the zone of maximum sea-surface temperature have been increased another 0.2-0.4°C. The positive sea surface temperature anomalies reach 1.5-2.0°C over the central and eastern Pacific. The sea surface temperatures along the South American coast have returned to near normal, the easterlies in the eastern equatorial Pacific are stronger than during the pre-onset phase. The magnitudes of low-level equatorial convergence over the central Pacific are increased over that of the transition phase by about 20 percent.

f. Applications of ENSO - Fraedrich (1988) used a 140-year (1844-1983) record of ENSO observations to show the predictability of ENSO. It correlates to within about one to two years, provided that the initial conditions are reliable. This means the long term predictions are still difficult. Gray (1984a,b) found that the yearly numbers of Atlantic hurricanes are correlated with the El Nino phenomena. Chan (1985) did a similar study for the Pacific. The ultimate causes of the El Nino effects are not yet satisfactorily established.

1.3.2 Tropical and Mid-latitude Interaction

1.3.2.1 General

Tropical weather forecasters must have a sound understanding of dynamics of both midlatitude and tropical circulations and must extend their analyses sufficiently poleward to identify and interpret tropical/mid-latitude interactions (Atkinson 1971).

Mid-latitude circulations often affect tropical circulation and weather patterns, and vice versa. Time-lapsed movies of cloud imageries from geostationary satellites often show cloud bands extending from the mid-latitudes into tropics. During winter, the cold fronts and cold surges can penetrate into the tropics and interact with weather systems in low latitude. During summer, tropical cyclones occasionally move northward and become extratropical cyclones.

1.3.2.2 Large-Scale Monsoon

The precise definition of a monsoon is a much debated issue. In general, the term monsoon means a change in persistent wind direction according to season. Winds flowing persistently from one direction during summer shift to a different (not necessarily opposite) direction during winter. Monsoons are important to tropical cyclone forecasting because they directly affect tropical cyclone motion and decay.

Large-scale monsoons can be viewed as three-dimensional circulations associated with the global distribution of land and water. The common elements in both summer and winter large-scale monsoons are:

a. Lower tropospheric winds flow from a large high-pressure area into a large low-pressure area. The Coriolis effect influences the large-scale winds.

b. The low-pressure area is usually formed as a belt-shaped zone, and is called the monsoon trough. Monsoon troughs are associated with a large area of cloudiness.

c. The wind speeds between the high-pressure area and the monsoon trough are usually concentrated into a jet. This is referred to as a monsoon jet or surge, a low-level jet, or a cross equatorial jet. Features associated with the jet can be complex and can have local characteristics.

d. The monsoonal circulation has a three-dimensional structure. The rising air along the monsoon trough diverges out of an upper-tropospheric anticyclone along an upper-level jet-like flow.

1.3.2.3 Monsoon Indices

Monsoon indications have been developed to determine which areas of the world actually experience monsoons. These indices typically overestimate the monsoonal areas, including such regions as the Russian Arctic and the Gulf of Alaska. Ramage (1971) listed four criteria which define a region as monsoonal:

a. the prevailing wind direction shifts by at least 120° between January and July

b. the average frequency of prevailing wind directions in January and July exceeds 40%

c. the mean resultant winds in at least one of the months exceed 3 ms^{-1}, and

d. fewer than one cyclone-anticyclone alternation occurs every two years in either month in a 5° latitude-longitude rectangle.

Using this definition, the monsoon regions are central Africa, India and the North Indian Ocean, North Australia, Indonesia, and Southeast Asia (Fig 2.24). Since the monsoon of Africa does not play a role in tropical cyclones, it will not be discussed here.

1.3.2.4 Indian-Asian Southwest Summer Monsoon

The Indian-Asian summer monsoon is the most widely studied. During the winter Fig 2.6), the temperature typically decreases poleward of the equator. With the heating of the continents during summer, especially the Tibetan plateau, the temperature gradient is reversed. The region of highest temperatures is now located near the southern base of the Himalayas. The winter northeasterly flow reverses to southwesterly, lasting from May to October. The surface wind chart for July (Fig 2.8) is typical of the monsoonal flow pattern. The trade wind southeasterlies of the South Indian Ocean turn southerly, then southwesterly north of the equator. A speed maximum, often referred to as the Somalia jet, exists in the Arabian Sea near the Gulf of Aden.

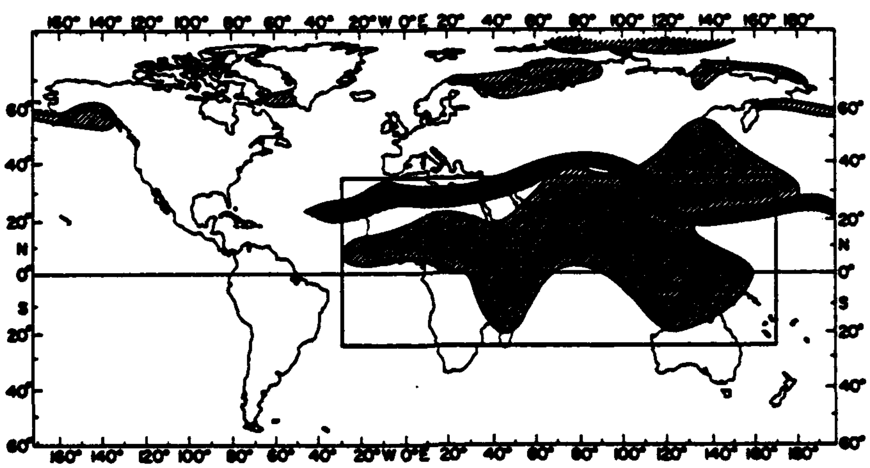

Fig. 2.24. Final delineation of monsoon regions. Hatched areas are "monsoonal" according to Khromov (1957). Heavy line marks northern limits of the region within the Northern Hemisphere with low frequencies of surface cyclone-anticyclone alterations in summer and winter (Klien, 1957). Rectangle encloses the monsoon region.

The influence of the Indian-Asian monsoon extends from eastern Africa well into the western North Pacific near 150E (see Fig. 2.4). As mentioned in the general circulation section, the monsoon trough of the western North Pacific is poorly depicted on a mean chart due to its migratory nature.

Aloft, a large anticyclone is situated over the Tibetan Plateau. The easterlies to the south serve as the return branch of the surface monsoonal flow. This results in large, persistent vertical shear, making tropical cyclogenesis very rare during these months. During the period 1975-1990, there were an average of 0.5 tropical cyclones in June, 0 in July, 0.1 in August, and 0.2 in September.

Of all of the phenomena associated with the Indian-Asian monsoon, the onset is probably the most interesting. The interest lies in the suddenness of the onset, which can typically be determined to the day using rainfall as the indicator. As indicated in the section on surface circulation, it typically occurs during May. But the actual date of the onset varies throughout the region. Table 2.1 gives the typical onset date for various locations.

Table 2.1 Statistics of onsets of Indian-Asian summer monsoon (Orgill, 1967).

Area	Mean Date	Earliest Date	Latest Date	Range (in days)
Southeast Asia	May 17	May 1	June 3	33
India, W. Coast (south)	May 31	May 11	June 6	30
India, W. Coast (north)	June 8	May 21	June 18	28
India, New Dehli	July 2	June 17	July 20	34

1.3.2.5 Southeast Asian Winter Monsoon

As with the southwest summer monsoon (Fig. 2.8), the winter monsoon (Fig 2.6) owes its existence in large part to the Tibetan plateau. The east-west orientation of the Himalayas blocks the synoptic scale exchanges of cold polar air with warm tropical air. The only avenue of exchange is east of the Himalayas, over southeast Asia. Consequently, cold air from the Siberian anticyclone flows southward across eastern China, over the South China Sea, and into Southeast Asia. Numerical modeling studies show that if the Himalayan mountains were "removed", the Siberian anticyclone would no longer be a semi-permanent feature. Rather, the region would resemble North America with transitory cyclones and anticyclones.

The onset of the northeast monsoon is far more distinct than the summer monsoon. Cold air begins to penetrate North Vietnam during late August and September. By the end of October, the entire Indochina Peninsula is covered by the winter northeasterlies. The monsoon typically lasts through March.

During the northeast monsoon season, the South China Sea region is characterized by strong vertical shear. The low-level monsoon northeasterlies are overlaid with strong westerlies from India. This environment makes tropical cyclogenesis rare. More importantly, tropical cyclones moving into this region are typically sheared, with the low-level circulation steered to the southwest.

As with the summer southwest monsoon, the winter monsoon, while persistent, experiences surges and lulls. Surges occur when an upper-level trough develops over northern China and moves off the east coast 24 hours later. Lulls are associated with surface bubble highs moving off the east coast of China, creating easterly or southeasterly flow over the South China Sea.

1.3.2.6 Northwest Australian Summer Monsoon

Like its Indian counterpart, the Australian summer monsoon (Fig. 2.2) is largely driven by the strong heat lows which form over Northern Australia. However, since Australia lacks a mountain range like the Tibetan Plateau, the monsoonal flow is neither as strong nor as steady as the Indian monsoon.

The onset of the monsoonal flow typically occurs during January. As the southeast Asian winter monsoon strengthens, its flow penetrates deeper and deeper into the tropical latitudes. By January, the flow crosses the equator near Indonesia and turns westerly. These equatorial westerlies flow across New Guinea, the Solomon Islands, and Northern Australia. The season lasts until April when easterly trades return (Fig 2.3), covering the entire area south of about 10S.

As with the Indian monsoon, a semi-permanent anticyclone is located over the surface heat lows. Upper-level southeasterlies to the north of this anticyclone cross the equator, providing the return flow of the Hadley circulation. This reversal in flow (upper-level southeasterlies over low-level northwesterlies) results in a large vertical shear similar to the

Indian summer monsoon. However, in this case, the shear region is only located equatorward of 10S. Thus, unlike the Indian monsoon season, tropical cyclogenesis still occurs during the Australian monsoon season, but is confined to south of 10S.

1.3.3 Interactions Between Troposphere and Stratosphere

A review of the mean zonal temperature shows that the tropical stratosphere, is thermally stable. Wallace (1973) reviewed the general circulation of the tropical lower stratosphere and discovered three periods of oscillation:

1.3.3 1 Quasi-biennial Oscillation (QBO)

The QBO in tropical stratospheric winds is defined as a reversal of the mean winds over a fixed location from easterlies to westerlies. It has been determined that this reversal occurs with mean periods around 26 months, and a range of 23 to 29 months (Fig. 2.25).

1.3.3.2 Annual Oscillation (AO)

The AO is defined as the tendency of the lower stratospheric winds to become easterlies in the summer hemisphere and westerlies in the winter hemisphere. The AO is primarily an extratropical phenomenon and does not interact strongly with tropical circulation systems.

1.3.3.3 Semiannual Oscillation (SAO)

The SAO of lower stratospheric zonal wind reaches its maximum amplitudes in extratropics and is an indicator of midwinter breakdowns in the polar night jet (Matsuno, 1970).

1.3.3.4 Applications of QBO

The QBO may be associated with the seasonal weather activities. Gray (1984a,b) has used the QBO at the 30-mb level as one of the indexes to predict the yearly number of tropical cyclones in the Atlantic with some success. However, the physical links between cyclone activity and QBO are not clearly understood.

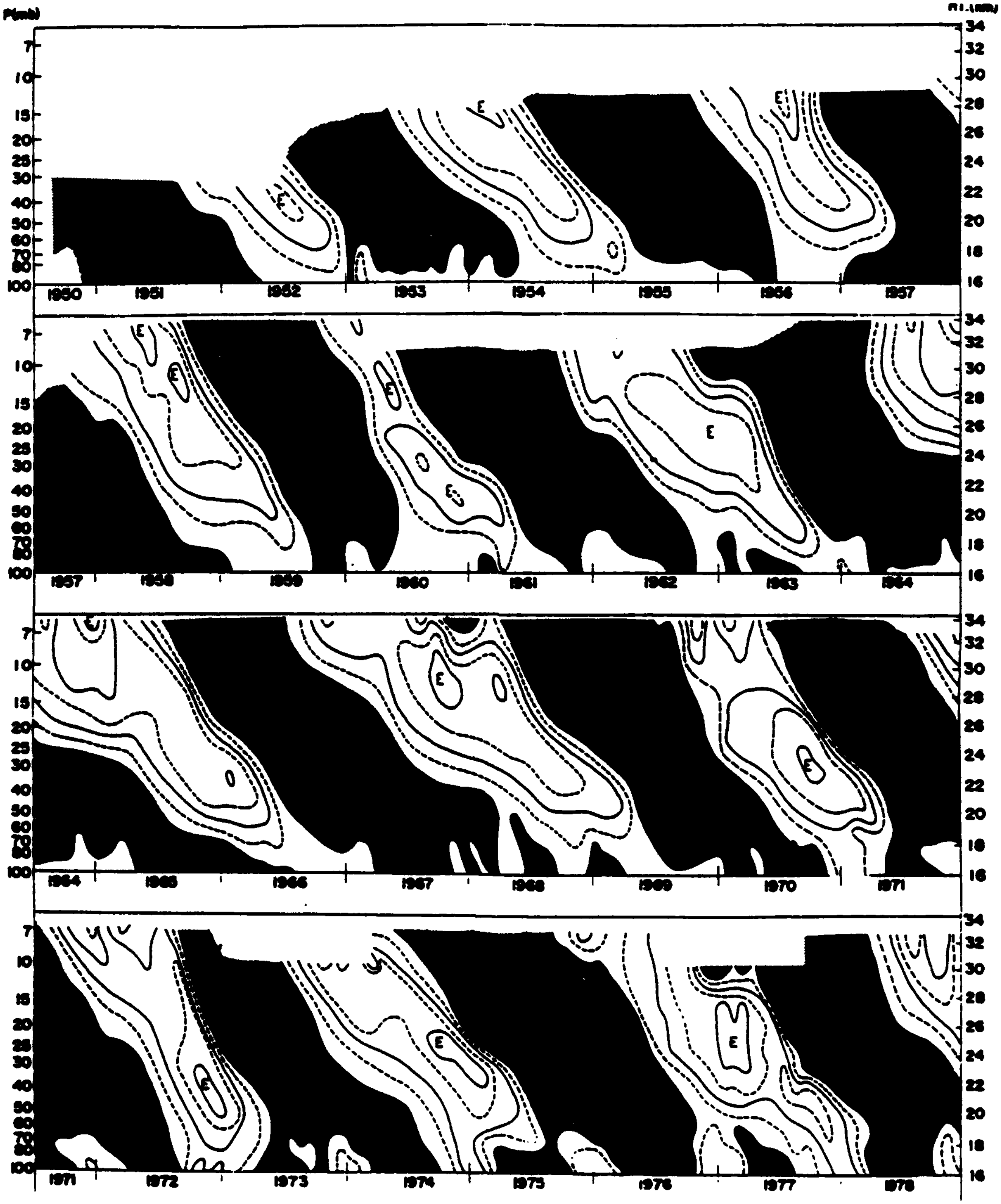

Fig. 2.25. Time-height section of the zonal wind near 9N with 15-year average of the monthly means subtracted to remove annual and semiannual cycles. Solid isotachs are placed at intervals of 10 m/s. Shaded areas indicate westerlies (Wallace, 1973).

2. SYNOPTIC MODELS AND TROPOSPHERIC WAVE DISTURBANCES

La Seur (1964) developed an excellent review of the synoptic models in the tropics. According to his classification, there are three types of tropical synoptic systems: waves, vortices and linear disturbances.

2.1 Waves

Several modes of tropical wave disturbances in the tropospheric easterlies have been studied extensively: the westward- moving North African easterly waves (Burpee,1972), the eastward-moving equatorial Africa waves (Johnson, 1964), the westward moving Eastern Atlantic easterly waves (Piersig,1936), the Caribbean easterly waves (Dunn,1940; Riehl,1945), and the Pacific equatorial waves (Palmer, 1952). These waves are usually observed in the zone between the equator and 30N.

Figures 2.26 and 2.27 (Nieuwolt, 1977) show a top view and cross section of an idealized easterly wave in the Caribbean. These waves move westward and usually last from one week to several weeks. The wave speeds are 10 to 15 knots (about 5 to 7° latitude per day) and are usually slower than the easterly current in which they are embedded. The wave pattern can be seen in the streamline charts, with maximum amplitude in the layer between 500 to 700 mb levels. The characteristic wavelength is of the order of 2,000 km. The characteristic periods of waves can vary from 2 to 5 days. The line of maximum wind shear has a somewhat north-to-south orientation. There is generally fair weather to the west of the trough line, where divergence, subsidence and drier air dominate the low troposphere. The moisture layer raises sharply near the trough, intense low tropospheric convergence and moisture, convective cloudiness, and rainfall prevail in and to the east of the trough. Cloud clusters, seen on satellite cloud imageries, are usually associated with the waves. A majority of these waves are cold-core, and the axis of the wave tilts toward the east with height. Some squalls and tropical cyclones are believed to develop from cloud clusters associated with the waves (Yanai,1961, 1968).

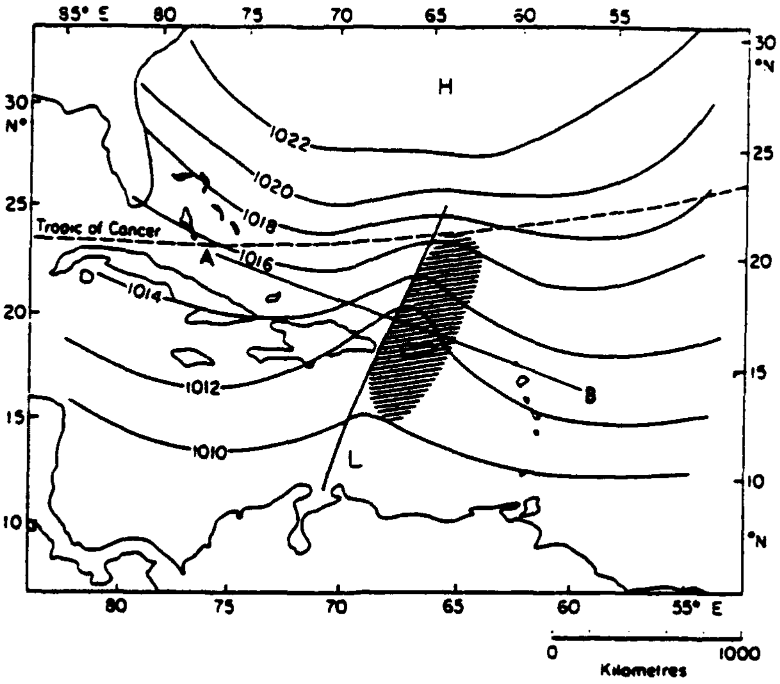

Fig. 2.26. Model of an easterly wave in the Caribbean area. Hatched area indicates main rainfall zone (Nieuwolt, 1977).

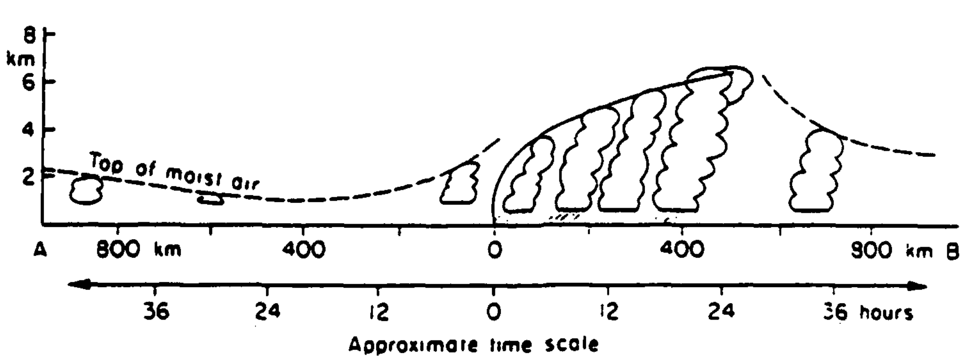

Fig. 2.27. Cross section along the line A-B in Fig. 2.26, showing the vertical structure of an easterly wave. Vertical exaggeration about 50 times (Nieuwolt, 1977).

The easterly waves on the daily weather map can have large deviation from the idealized model discussed above. Riehl (1967) studied a wave in the western Caribbean and found the wave was moving westward faster than the basic current. The convergence area was located at the trough. Frank (1969) found an inverted-V shaped cloud pattern associated with Caribbean easterly waves. These waves move westward at an average speed of about 16 knots (8° latitude per day). Riehl (1979) pointed out that the basic easterlies in a region may differ substantially from year to year. Therefore, there will be many waves in some years, such as in the Caribbean in 1969, and few waves in other years, such as 1972.

Atkinson (1971) offered a warning that the easterly-wave models mentioned above have been grossly overworked and misused, especially as an aid to tropical forecasting. It is now known that there are a wide variety of weather-producing systems. Some easterly waves in the lower tropospheric easterlies are the reflections of upper tropospheric disturbances (Yanai and Nitta, 1967; Yanai, et al., 1968).

2.2 Synoptic Scale Vortices

A *synoptic* scale vortex usually occupies an area with a length of about 300 km to 1000 km. La Seur (1964) classified tropospheric vortices into two types; one has maximum intensity in the lower troposphere, the other has maximum intensity in the upper troposphere.

2.2.1 Lower Tropospheric Cyclonic Vortices

The lower tropospheric cyclone ranges in intensity from barely detectable cyclones, to super typhoons. They originate from synoptic scale disturbances associated with shear zones (or areas of low-level convergence/confluence) in the tropics. The release of latent heat from condensation plays a significant role in the formation and maintenance of these systems. Their movement within the tropics is predominantly westward. Figure 2.28 (Atkinson and Sadler, 1970) shows the four monthly mean locations of the monsoon troughs, trade wind troughs and equatorial buffer zones/near equatorial troughs. The trade wind trough in the central Pacific is difficult to locate in the resultant gradient-level wind field during some months so it has been omitted in those areas.

Sadler (1967) labeled the areas of tropical-cyclone origin with capital letters in Fig. 2.28; the characteristic of each area are discussed in the following:

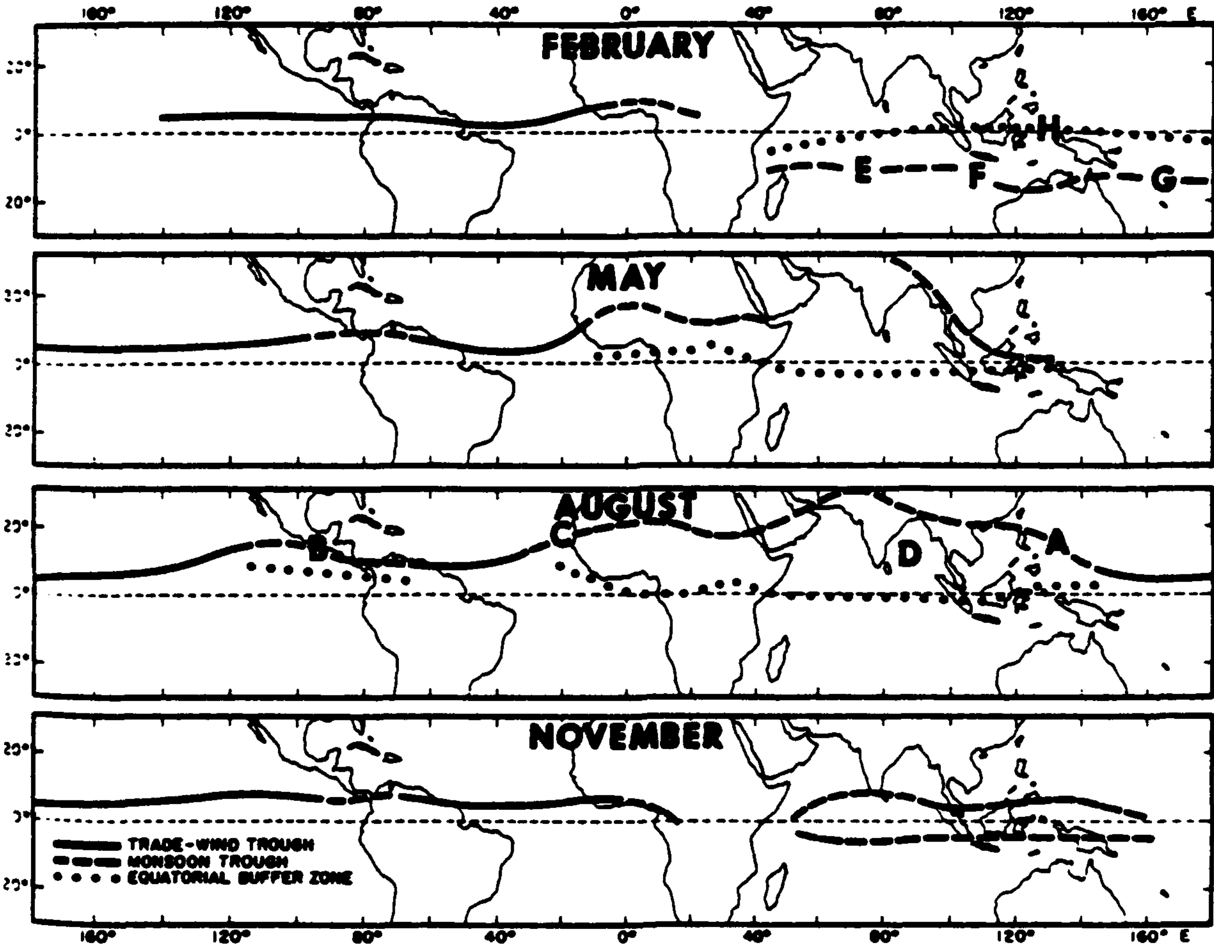

Fig. 2.28. Locations of the trade wind troughs, monsoon troughs, and equal buffer zones at the gradient level during February, May, August, and November (Atkinson and Sadler, 1970).

2.2.1.1 Western North Pacific Summer - Area A

This area has maximum frequency of tropical cyclones. The monsoon trough merges into a trade wind trough to the east. The frequency of cyclone formation appears to be greatest when the location of the monsoon trough is north or extends eastward of its mean location.

2.2.1.2 Eastern North Pacific - Area B

The monsoon trough merges into a trade wind trough to the west. This area has the secondary peak frequency of tropical cyclone development. The cyclones, once formed, usually move westward or northwestward into a region of cooler sea-surface temperature and stronger vertical wind shear and eventually dissipate.

2.2.1.3 Eastern North Atlantic - Area C

The cyclones in this area originate over land in the shear zone over Africa but not at the surface level. The shear zone slopes equatorward from a mean location of 20N to 25N near the surface to 10N to 15N at the 700 mb level. There is extremely dry air at the surface associated with the shear zone. A deep moist layer is located to the south of the surface shear zone. Vortex and wave disturbance are frequently detected in this deep moist layer (see Section 2.1) . This results in cyclones developing over Africa that are most intense near 5,000 to 10,000 feet (800 mb to 700 mb level). These cyclones usually do not have an associated surface-wind vortex until they reach the West coast of Africa. Carlson (1969a) studied three disturbances moving across Africa during August and September 1967. He concluded that the cyclonic vortices are best tracked at the 700 mb level. The westward speed of movement of the vortices are about 12 to 20 knots (5° to 9° latitude per day). As the vortices move across the North Atlantic, some intensify into tropical storms and hurricanes. This intensification can occur from the mid-Atlantic all the way throughout the Caribbean.

2.2.1.4 North Bay of Bengal - Area D

The monsoon trough in this area is normally located over land during the southwest monsoon season from June to September. The monsoon trough occasionally moves southward into the Bay of Bengal, and tropical depression can develop in the vicinity of the trough. These depressions usually move westward or northwestward across India along the monsoon trough toward the northwest. These depressions occur about twice a month during the monsoon season, and are called monsoon depressions (Sikka, 1977). Monsoon rainfall is highly variable from year to year. Breaks occur within the monsoon when the jet stream or large-amplitude troughs from midlatitude systems move southwards weakening the Tibetan anticyclone (Ramaswamy, 1962). During the months of April, May, October, November and December, vortices form further south, in the monsoon trough, and often reach tropical storm or typhoon intensity before striking land.

2.2.1.5 Southern Hemisphere - Areas E, F, and G

In this area, the vortices can form in the monsoon trough which extends eastward from Africa to near 180°. Vortex formation usually occurs during the period from November to May and is most frequent from January to March. Vortices generally do not occur in the eastern South Atlantic and South Pacific where monsoon troughs do not exist.

2.2.1.6 Western North Pacific Winter - Area H

Tropical cyclones can develop in this area during the winter season when the low tropospheric wind-shear zone is located near 5N. The majority of these cyclones do not develop beyond tropical depression stage, but a few do become tropical storms and typhoons.

2.2.2 Low Tropospheric Anticyclonic Vortices

2.2.2.1 Low-level Anticyclones

These occur in regions where the ITCZ or equatorial trough is displaced more than about 10 degrees of latitude from the equator. They weaken with height, and are not seen above 20,000 feet (about 6 km) and move predominantly westward. They are typically associated with above average convection and cloudiness in the poleward sectors.

2.2.2.2 Fujita et al. (1969) Equatorial Anticyclone Model

This study combined geosynchronous satellite cloud-track winds, sounding winds, and corresponding vorticity and divergence fields, to develop a conceptual model of westward moving equatorial anticyclones in the eastern North Pacific during the warm season. Figure 2.29 shows the six stages of the Fujita et al. model:

a. Pushing Stage - The low-tropospheric large-scale flow from the Southern Hemisphere pushes northward, thus producing a shear/low-level convergence zone and an associated intertropical cloud band. During this stage the center line of the cloud band can be pushed to 10N. Formations of tropical depressions are often observed along the zone in which Southern and Northern Hemisphere air begins interacting with large horizontal wind shear and cyclonic relative vorticity.

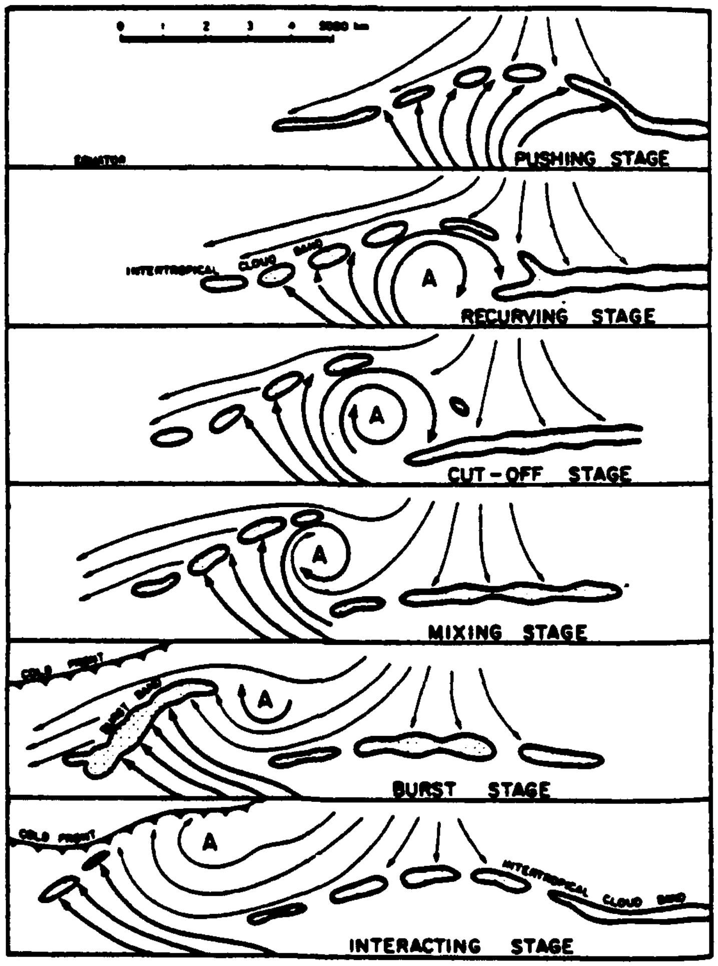

Fig. 2.29. A proposed model of an equatorial anticyclone in six stages. (Fujita et al., 1969).

b. Recurving Stage - The flow from the Southern Hemisphere gains sufficient anticyclonic relative vorticity to start the flow returning southward. Tropical depressions formed during the pushing stage tend to move out of the intertropical cloud band.

c. Cut-off Stage - After a day or so, the equatorial anticyclone will become characterized by closed or almost closed circulation. The anticyclone center is encircled entirely by the air from the Southern Hemisphere. The intertropical cloud band starts to break into two bands, one located to the northwest of the anticyclone center, the other located to the east of the center.

d. Mixing Stage - Following the break in the intertropical cloud band, the Northern Hemisphere trade winds start flowing to the south of the anticyclone center. The Northern Hemisphere trade winds and the flow from the Southern Hemisphere start mixing around the anticyclone.

e. Burst Stage - After a day or so, a significant amount of the northern trade winds have been transported through the southern sectors of the anticyclone. These northern trade winds begin to migrate toward the west-northwest. Meanwhile the flow from the Southern Hemisphere keeps pushing the cloud band along the leading edge of anticyclone. Such a joint push by both northern trade winds and the flow from the Southern Hemisphere often result in an intense zone of convergence with cyclonic vorticity. The cloud band located in this zone is called a burst band. This stage lasts only one or two days and then the burst band disintegrates into small fragments.

f. Interacting Stage - The southerly or southeasterly flow to the south of the equatorial anticyclone center becomes stronger. The anticyclonic flow is still strong enough to resist the southeastward movement of a mid-latitude cold front. The anticyclonic center tends to move northwestward.

2.2.3 Middle Tropospheric Cyclonic Vortices

Atkinson (1971) classified the mid-tropospheric cyclones into the subtropical cyclones and the Arabian Sea cyclones. These two classes are discussed in detail.

2.2.3.1 Subtropical Cyclones or Kona Lows

Palmen (1949) described the sequence of events leading to a cut-off low equatorward of the polar jet. These cyclones usually occur during winter in the eastern North Pacific and North Atlantic. A cyclone frequently develops leeward of the prevailing northeast trade winds and is called kona in Polynesian. These kona cyclones are usually associated with southeastward movement of the Aleutian low and the passage of a secondary depression from northwest to southeast, north of the Hawaiian Islands. Simpson (1952) classified cyclones into two groups; occluded-cyclone source and cyclogenesis source.

a. Occluded-Cyclone Source - This type of low develops when a fully mature occluded cyclones moves southeastward into the central Pacific from the Asiatic mainland and maintains very pronounced cold fronts. The continental polar air mass or migratory anticyclones associated with these cold fronts move into the eastern Pacific. As they continue across the Pacific they frequently encounter a semi-permanent anticyclone which carries tropical air northward into contact with the north wind of the migratory anticyclone developing a stationary front or shear zone. Frontal waves form in this area and tend to become unstable and rapidly occlude. The migratory anticyclones continue to move rapidly eastward with speeds about 20° latitude per day and merge with the semi-permanent anticyclone. This configuration of circulations can trap the occluded cyclone and cut off the cold-air supply. In this declining stage of the life cycle of this misplaced extratropical cyclone, the kona cyclone is born. The corresponding circulation in the middle and upper troposphere usually develops into a closed low that is cut off from the main westerlies. In a 20-year observation period 76 of these cyclones developed.

b. Cyclogenesis Source - The second cyclogenesis source describes a situation when cut-off cold-core lows in the upper troposphere extend their circulation to the low-levels in the subtropical eastern Pacific during the winter season. These cyclones develop on the eastern side of an omega blocking high and move southwestward for several thousand kilometers before they dissipate or merge with flow from a deepening trough in the middle latitude westerlies. During most of the year, subtropical cyclones in the eastern Pacific do not extend their circulations below the 600 mb level. Even in these cases, their approach and passage over Hawaii has a marked effect on rainfall. During the winter season, from November to March, the middle tropospheric cyclones occasionally penetrate to the surface, causing cyclogenesis in the easterlies.

c. Atlantic Kona-Type Cyclones - In the Atlantic, kona-type cyclones are observed to form southwest of the Azores and move mainly toward the south and the west.

d. Subtropical Cyclone Wind/Rainfall Distribution - Figures 2.30 and 2.31 shows the surface wind and rainfall distribution of a composite Kona cyclone. The areas of maximum wind speed and rainfall are located east of the circulation center, while secondary peaks of wind speed and rainfall are located to the northwest. The center is an area of light wind speeds and minimal rainfall.

2.2.3.2 Arabian Sea Cyclones

These mid-tropospheric cyclones are often observed on daily and monthly charts over the west coast of India during the southwest monsoon season. The cyclones develop between the 700 to 500 mb levels in the monsoon trough (Fig. 2.32). The monsoon trough generally

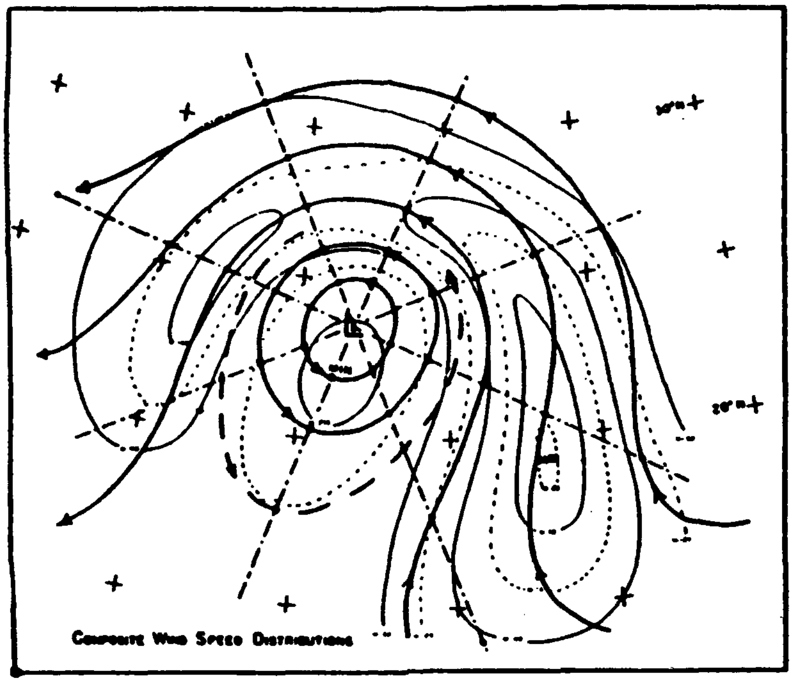

Fig. 2.30. Composite chart of wind-speed distributions in a Kona storm. Units are miles of wind movement in 24 hours, expressed in terms of deviation from maximum.

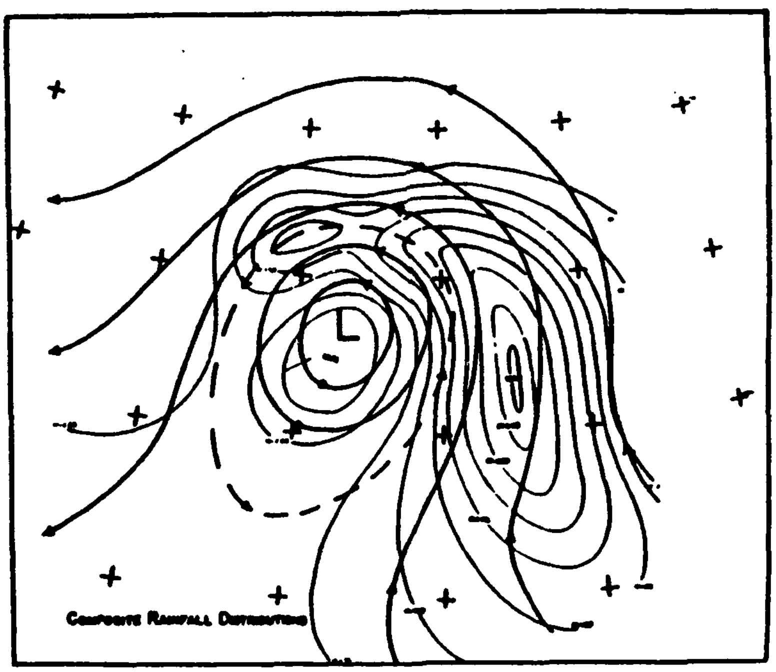

Fig. 2.31. Composite chart of rainfall distributions in a Kona storm. Units are inches of rain per 24 hours, expressed in terms of deviation from maximim.

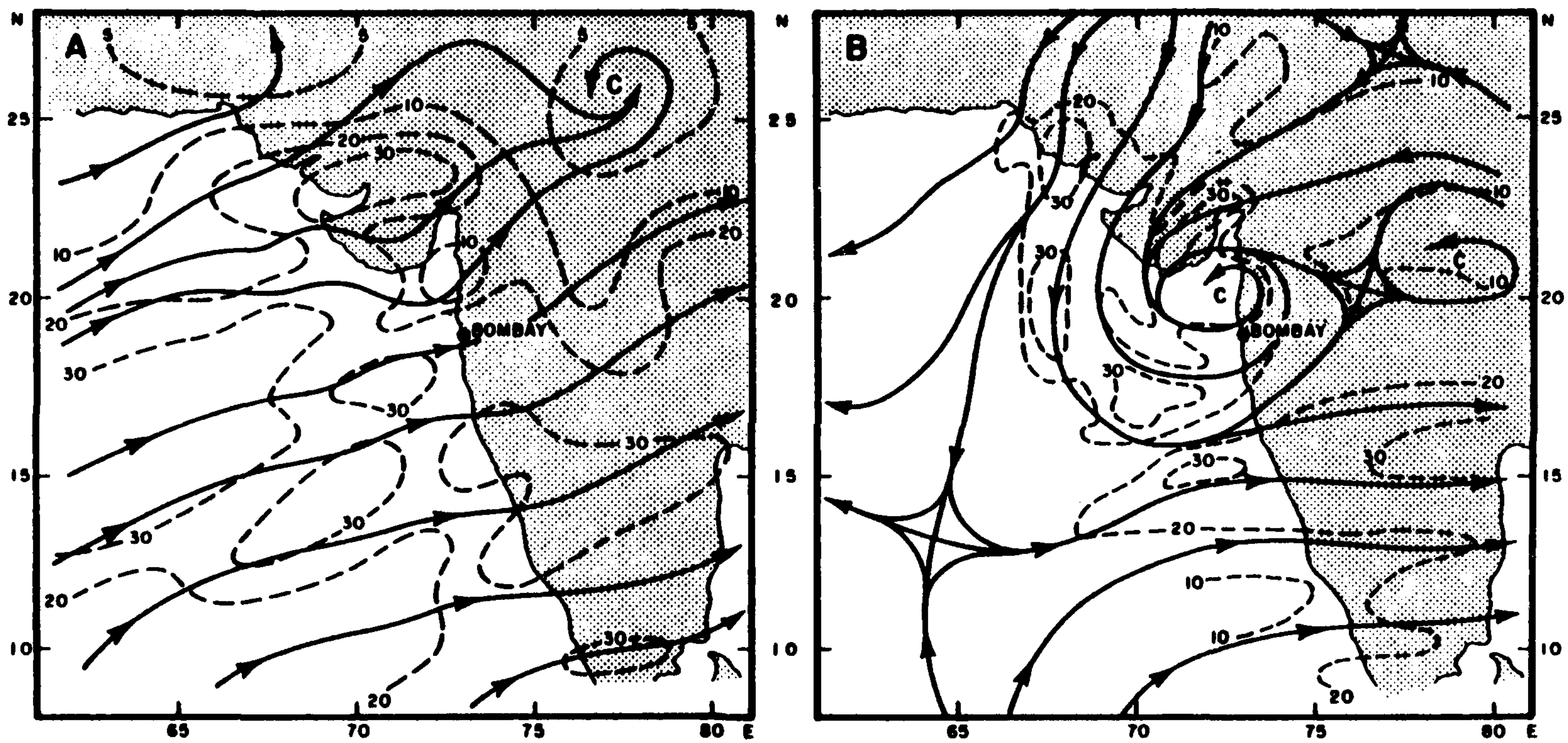

Fig. 2.32. Composite kinematic analyses for (A) a near-surface layer (500-900 meters), and (B) the 600 mb level showing a well-veloped mid-tropospheric cyclone over western India during July 1963 (Atkinson, 1971).

tilts equatorward with height, and vanishes above 400 mbs, where a steady easterly regime prevails. The portion of the monsoon trough in the Arabian Sea has more vertical tilt than that along the east coast of India. Surface cyclones occur frequently over the east coast of India and Bay of Bengal but rarely are observed over the west coast of India and northeast Arabian Sea. In the Arabian Sea, the mid-tropospheric cyclones are the dominant weather phenomena. Fig. 2.32 shows a cyclone near Bombay during the period 1-10 July 1963. The cyclonic signature at the 600 mb level is much stronger than that at the surface. Maximum convection is located west of the cyclone center.

2.3 Upper tropospheric Vortices

2.3.1 General Characteristics

The upper tropospheric vortices usually do not extend below 20,000 feet (about 6 km). A weak inverted wave in the easterlies is usually found beneath these vortices. They may be associated with broad areas of high clouds. Downward growth of an upper-level vortex results in the appearance of a surface vortex and increases in convective clouds. On rare occasions the vortices become warm-core and develop into tropical cyclones. Vortex movement is usually slowly west-southwestward.

2.3.2 Documented Findings

Sadler (1964) documented the tropical cyclones of the eastern North Pacific as revealed by satellite observations, and proposed that the upper tropospheric circulation is a controlling factor in the development and life history of hurricanes. He suggested that the regions of hurricane development in the eastern Pacific and typhoon development in the western Pacific have similar locations relative to the mean upper tropospheric trough and the westward ends of the ridge lines. This kind of trough is now referred to as tropical upper tropospheric trough or TUTT.

2.3.3 Upper Tropospheric Trough Characteristics

Atkinson (1971) reviewed the characteristics of upper tropospheric cyclones in the following four regions:

a. North Pacific - In the western North Pacific, there are high correlations between the locations of formative tropical cyclones and that of the lower tropospheric monsoon troughs and the TUTT. The TUTT usually occurs from May to November, and is most active between July and September. Sadler (1976a) proposed a revised synoptic model for the TUTT during the early typhoon season (i.e., June-August) in the western Pacific, this model will be discussed in detail in the Genesis Section. Sadler (1976a) suggested that the TUTT cyclone cells originate from the dynamic coupling between the TUTT and transitory midlatitude disturbances when the subtropical high between these systems is unusually weak.

b. North Atlantic - The upper tropospheric trough is used to describe the major semipermanent circulation pattern that develops in the North Atlantic during the period form August to November. Carlson (1967) analyzed synoptic data over the eastern Caribbean for October 1965 and identified the presence of an upper tropospheric cold core cyclone. These cold core cyclones usually originate near the Azores and move westward along the latitude 20N. The systems extend over an area about 20 degrees latitude and 40 degrees longitude. The lowest level of a closed circulation associated with the upper cold core cyclone is commonly from the 700 to 500 mb level. This feature in the North Atlantic is different from that in the North Pacific. Most of the systems are detectable in the low tropospheric temperature field as cold troughs in the easterlies. The cyclones tend to tilt vertically toward the northeast. Convective clouds and rainfall occur in the southeast quadrant, about 5 degree latitude from the upper cyclone center. Large variations of cloudiness can exist from one system to the next.

Sadler (1967a) pointed out that the summer TUTT is a dominant feature over the trade wind regions of the North Atlantic Ocean, the Gulf of Mexico and the Caribbean Sea. He also noted that lower tropospheric responses to the TUTT in the North Atlantic are different than those in the North Pacific.

c. South Pacific - Tropical cyclones develop during the warm season in the areas below the tropical upper tropospheric trough over the South Pacific. The mean location of the trough is in the area from the equator to 30S and from 180E to 120E. Sadler (1976a) stated that the summer TUTT in the Southern Hemisphere lies over the trade wind region of the east central Pacific and can induce tropical cyclones.

2.2.4 Upper-tropospheric Anticyclonic Vortices

Upper-level anticyclones occur in the summer hemisphere and are seldom found below 15,000 feet (about 500 mb level). They have warm-core structures. According to the cloud distribution, there are two types of warm-core structures; one is accompanied by little or no high cloudiness and is due to large-scale subsidence motion. The other is accompanied by convective clouds and is due to cumulus-scale release of latent heat.

2.3 Linear Disturbances

In general, air masses in the tropics are much more homogeneous than those in the midlatitudes. It is therefore almost impossible to define a surface front in the tropics using the polar front theory. Since oceanic tropical air is usually conditionally unstable, cyclonic shear lines or convergence lines are frequently associated with organized deep convection. These linear disturbances are also called squall lines or sea breeze fronts. Two well studied tropical linear disturbances are the squall lines of West Africa and the sumatras of Malaysia.

3. TROPICAL CYCLONES

Severe weather, strong gusts and heavy rain, and high seas associated with tropical cyclones are critical parameters affecting the Navy's operations. Calhoun (1981) documented the damages to the Third Fleet due to a Pacific storm. The methods of operational forecasts for onset, motion and decay of tropical cyclones are key subjective decisions made based on various objective aids in a man-machine mix environment.

3.1. Classifications of Tropical Cyclones

Tropical cyclone is a general term for a cyclone that originates over the tropical oceans. The tropical cyclone is a nearly circular storm with a mean diameter of about 5 degrees latitude and an extremely low pressure center into which the horizontal winds spiral. The life span of a tropical cyclone is about 5 to 10 days. Table 2.2 is a classification of tropical cyclones according to their intensity, which is the maximum wind speed near the surface (Huschke, 1959):

Table 2.2. Tropical Cyclone Classification

World Meteorological Organization	United States Of America	Japan (international code)
	Tropical Disturbance slight surface circulation with one closed surface isobar or none	
Tropical Depression winds up to 34 knots	**Tropical Depression** winds up to Beaufort Scale 7 (28 knots)	**Tropical Depression** winds up to Beaufort Scale 7
Tropical Storm 35 - 40 knots	**Tropical Storm** Beaufort 8 - 11 (34 - 63 knots)	**Tropical Storm** **8 - 9 (34-47 knots)** - - - - - - - - - - - - - - - - - **Severe Tropical Storm** 10 - 11 (48-63 knots)
Hurricane or Typhoon 65 knots or higher	**Hurricane** Beaufort 12 or more (64 knots or more)	**Typhoon** Beaufort 12 or more

One exception to the table is that India uses the terms Moderate Tropical Storm and Hurricane, instead of the terms Tropical Storm and Typhoon.

Mature tropical cyclones (i.e., typhoons or hurricanes), range in size from 100 km to well over 1500 km in diameter. The distribution of tangential winds, pressure and temperature tends to be radially symmetric. However, asymmetric components are frequently detected by radar precipitation reflections and satellite cloud images. Strong wind speeds are present in most cases when the central pressure of the cyclone falls below 990 mb level, although there is no exact balance between the low-level mass and velocity fields.

3.2 Tropical Cyclone Formation Requirements

Current thinking identifies six prerequisites for tropical cyclone development. In general, all six of these parameters are satisfied over the tropical oceans at any given time. A significant deviation from the mean must occur with one or more of these variables for the formation of a tropical cyclone to occur. The following is a list of these parameters. For more detailed discussion see the Tropical Cyclone Formation chapter.

a. A Minimum Value of Earth's Vorticity - It has been observed that tropical cyclones do not form within 3 degrees of the equator. Apparently a minimum Coriolis effect must exist for a tropical cyclone to form.

b. Low-Level Relative Vorticity - Tropical cloud clusters which develop into tropical depressions are always located in regions of low-level positive vorticity (cyclonic turning).

c. Vertical Wind Shear - Large values of vertical wind shear prevent the formation of tropical cyclones.

d. SST and Mixed Layer Depth - Numerous studies have used exact numbers as a minimum SST criterion for development. Typically these are in the range of 26-27 degrees C. The warm ocean water must exist over a sufficient depth (i.e., 200 feet).

e. Potentially Unstable Atmosphere - The column must be potentially unstable to sustain convection for an extended period of time.

f. Mid-troposphere humidity: More vigorous convection occurs if the parcel remains saturated. A sufficient value of humidity is typically 50- 60% at lower to mid-levels.

3.3 Global Distributions of Tropical Cyclones

The global distribution is summarized in Table 2.3.

Table 2.3. Global Distribution of Tropical Cyclones.

Region	Season	Local Name
Western North Pacific South and East China Seas, Japan, Korea, Philippine	May - Dec	Typhoon, Tai-fung, Taifu Baguio or Baruio
Eastern North Pacific off the west coast of Mexico	May - Nov	Hurricane
Western North Atlantic Caribbean Sea and the Gulf of Mexico	Jun - Oct	Hurricane
Bay of Bengal	Apr - Dec	Cyclone
Arabian Sea	Jun - Oct	Cyclone
South Pacific off the coast of Australia and north of New Zealand	Dec - Apr	Cyclone
South Indian Ocean to the east of Madagascar	Nov - Apr	Cyclone
South Indian Ocean Off the northwest coast of Australia	Nov - May	Cyclone

The monthly tropical cyclone distribution is summarized in Table 2.4 (note the various periods of the data set), the number of years of data is in parentheses, the unit is number per year, a single 0 indicates no cyclone was recorded.

Table 2.4 Monthly Distribution of Tropical Cyclones.

	Months												
	1	2	3	4	5	6	7	8	9	10	11	12	SUM
Western North Pacific (30)	0.5	0.3	0.5	0.7	1.0	1.8	4.1	5.3	5.0	4.1	2.6	1.3	27.6
Eastern North Pacific (8)	0	0	0	0	4	1.5	3.6	4.1	3.0	1.9	0.4	0	14.9
Central North Pacific (22) (180-140 W)	0.0	0	0	0	0	0.2	0.9	1.6	0.8	0.3	0.1	0	3.9
North Atlantic (34)	0	0.0	0.0	0.0	0.2	0.6	0.8	2.4	3.5	1.8	0.4	0.1	9.8
North Indian Ocean (61):													
Bay of Bengal	0.1	0.0	0.1	0.2	0.4	0.5	0.6	0.4	0.5	0.9	0.9	0.4	5.0
Arabian Sea	0.0	0	0	0.1	0.2	0.2	0.0	0.0	0.1	0.3	0.2	0.1	1.2
Australian regions (10):													
North Coast	0.2	0.2	0.9	0.3	0	0	0	0	0	0	0.0	0.4	2.0
East Coast	0.9	0.8	0.9	0.3	0	0	0	0	0	0	0	0.2	3.1
West Coast	0.5	0.6	0.6	0.1	0	0	0	0	0	0	0	0.3	2.1
South Pacific (150 E-150 W) (40)	1.9	2.2	1.7	0.9	0.0	0.0	0	0	0	0.0	0.4	1.1	8.2

Data Sources:
Western North Pacific: Tachi, 1961; Arakawa, 1963; JTWC, 1989
Eastern North Pacific: Baum, 1974.
Central North Pacific: Shaw, 1981
North Atlantic: Dunn, 1956; Dunn and Miller, 1964; Nuemann et al., 1978.
North Indian Ocean: Koteswaran, 1963; Das, 1968.
South Indian Ocean: Ramage 1974.
Australian Region: Brunt and Hogen, 1956; +Lourensz, 1981.
South Pacific: Gabites, 1956, 1963; Revell, 1981.

Note: When more than one climatology value was available, the lowest value is reported.

REFERENCES

Arakawa, H., 1963: Typhoon climatology as revealed by data of Japanese Weather Service. Proc. Inter-regional Seminar on Tropical Cyclones in Tokyo, 1962. Japan Meteor. Agency, 31-36.

Aspliden, C. I., G. A. Dean, and H. Landers, 1965: Satellite study, tropical North Atlantic. Six reports date from 1965 to 1967. Dept. of Meteor. , Florida State Univ.

Atkinson, G. D., 1971: Forecasters' guide to tropical meteorology. US Air Force.

_____________, and J. C. Sadler, 1970: Mean cloudiness and gradient-level-wind charts over the tropics,. AWS Technical Report 215, Vols 1 and 2. US Air Force.

Barrett, E. C., 1974: <u>Climatology from satellites</u>. Methuen & Co, Ltd., London, England. 418 pp.

___________, and D. W. Martin, 1981: <u>The use of satellite data in rainfall monitoring</u>. Academic Press. 340 pp.

Baum, R. A., 1970: The eastern Pacific hurricane season 1973. <u>Mon. Wea. Rev</u>., 98, 280-292.

Bergeron, T., 1954: The problem of tropical hurricanes. <u>Quart. J. Royal Meteor. Soc</u>., 80, 131-164.

Bjerknes, J., 1959: The recent warming of the North Atlantic. In Rossby Memorial Volume, edited by Bolin, B. Rockfeller Inst. Press in association with Oxford Univ. Press. 65-73.

___________, 1961: El Nino study based on analysis of ocean surface temperature 1935-1957. Inter-American Tropical Tuna Comm. Bull., 5, 219-307.

___________, 1966: A possible response of the atmospheric Hadley circulation to equatorial anomalies of ocean temperature. <u>Tellus</u>, 18, 820-829.

___________, 1969: Atmospheric teleconnections from the equatorial Pacific. <u>Mon. Wea. Rev</u>., 97, 163-72.

Brunt, A. T. and J. Hogan, 1956: The occurrence of tropical cyclones in the Australian Region. Proc. Tropical Cyclones Sypm., Brisbane, Australian Bureau of Meteor, Melbourne. 5-18.

Burpee, R., 1972: The origin and structure of easterly waves in the lower troposphere of North Africa. <u>J. Atmos. Sci</u>., 29, 77- 90.

Calhoun, C. R., 1981: Typhoon, the other enemy. Naval Instit. Press. 247 pp.

Carlson, T. N, 1967: Structure of a steady-state cold low. Mon. Wea. Rev., 95, 763-777.

__________, 1969a: Synoptic histories of three African disturbances that developed into Atlantic hurricanes. Mon. Wea. Rev., 97, 256-276.

__________, 1969b: Some remarks on African disturbances and their progress over the tropical Atlantic. Mon. Wea. Rev., 97, 716-726.

Carr, F. H., 1977: Mid-tropospheric cyclones of the summer monsoon. Pageoph, 115, 1383-1412.

Chan, J. C. L., 1985: Tropical cyclone activity in the northwest Pacific in relation to the El Nino/southern oscillation phenomenon. Mon. Wea. Rev., 113, 599-606.

Chang, C.-P., 1970: Westward propagating cloud patterns in the tropical Pacific as seen from time-composite satellite photographs. J. Atmos. Sci., 27, 133-138.

Chang, J.H., 1972: Atmospheric circulation systems and climates. Oriental Publish Co. , Hawaii. 328 pp.

Charney, J. G., and M. E. Stern: 1962: On the stability of internal baroclinic jets in a rotating atmosphere. J. Atmos. Sci., 19, 159-172.

__________, and A. Eliassen, 1964: On the growth of the hurricane depression. J. Atmos. Sci., 21, 68-75.

Defant, F., 1951: Local winds. Compendium of Meteorology. American Meteor. Soc. 1334 pp.

Coy, L., 1980: Corrigendum to the manuscript "An unusually large westerly amplitude of the quasi-biennial oscillation. J. Atmos. Sci., 37, 912-913.

Das, P. K., 1968: The monsoons. National Book Trust, New Delhi, India.

Dobryshman, E. M., 1968: Some features of pressure and wind fields in the near-equatorial belt. Israel Program for Scientific Translations, Jerusalem. 205-221.

Dunn, G. E., 1940: Cyclogenesis in the tropical Atlantic. Bull. Amer. Meteor. Soc., 21, 215-229.

__________, 1956: Areas of hurricane development. Mon. Wea. Rev., 84, 47-51.

__________, and B. I. Miller, 1964: Atlantic Hurricanes. Louisiana State University Press. 337 pp.

Ebdon, R. A., 1960: Notes on the wind flow at 50 mb in tropical and subtropical regions in January 1957 and January 1958. Quart. J. Royal Met. Soc., 86, 540-542.

Fairbridge, R. W., Editor, 1967: The encyclopedia of atmospheric sciences and astrogeology. Reinhold Publishing Corporation, New York.

Flank, N. L., 1969: The inverted-V cloud pattern, and easterly wave? Mon. Wea. Rev., 97, 130-140.

Flohn, H., 1957: Large-scale aspects of the "summer monsoon" in south and east Asia. J. Meteor. Soc. Japan. Special 75th Ann. Vol., 180-186.

Fujita, T. T., K. Watanbe and T. Izawa: Formation and structure of equatorial anticyclones caused by large-scale cross-equatorial flows determined by ATS-1 photographs. J. Appl. Met., 8, 649- 667.

Gabites, J. F., 1956: A survey of tropical cyclones in the south Pacific. Proc. Tropical Cyclone Symp., Brisbane, Australian Bureau of Meteor., Melbourne. 19-24.

___________, 1963: The origin of tropical cyclones. Proc. Inter-regional Seminars on Tropical Cyclones in Tokyo, 1962. Japan Meteor. Agency. 53-58.

Gadgil, S., 1977: Orographic effects on the southwest monsoon: A review. Pageoph, 115, 1413- 1430.

Gibbs, W. J., 1956: Two cases of cyclogenesis in tropical latitudes. Proc. Tropical Cyclone Symp., Brisbane, Australian Bureau of Meteor., Melbourne. 275-288.

Gill, A. E., 1982: Atmosphere-ocean dynamics. Academic Press. 662 pp.

Gray, W. M., 1984a: Atlantic seasonal hurricane frequency. Part 1: El Nino and 30 mb QBO influences. Mon. Wea. Rev., 112, 1649- 1668.

___________, 1984b: Atlantic seasonal hurricane frequency. Part 2: forecasting its variability. Mon. Wea. Rev., 112, 1669-1683.

Hastenrath, S., 1988: Climate and circulation of the tropics. Reidel Publishing Co. 455 pp.

He H., J. W. McGinnis, Z. Song and M. Yanai, 1987: Onset of the Asian summer monsoon in 1979 and the effect of the Tibetan Plateau. Mon. Wea. Rev. 115, 1966-1995.

Hickey, J. R., L. L. Stowe, H. Jacobowitz, P. Pellegrino, R. H. Maschoff, F. House and T. VonderHaar, 1980: Initial solar irradiance determinations from Nimbus 7 cavity radiometer measurements. Science, 208, 281-283.

Huschke, R. E., 1957: Glossary of Meteorology. Amer. Meteor. Soc., Boston.

Johnson, D. H., 1964: Weather systems of west and central Africa. Proc. Symposium on Tropical Meteorology. New Zealand Meteor. Service, Wellington, New Zealand.

Johnson, H. M., 1976: A December cold front at the Hawaiian Islands. NWS/NESS Satellite Applications Information Note 1/76- 1, NOAA.

Jordan, C. L., 1955: Some features of the rainfall at Guam. Bull. Amer. Meteor. Soc, 36, 446-455.

JTWC, 1989: Annual tropical cyclone report. U.S. Naval Ocean. Command Center, JTWC, COMNAVMARIANAS Box 17, FPO San Francisco, CA 96630.

Kornfield, J., A. F. Hasler, K. J. Hanson and V. E. Suomi, 1967: Photographic cloud climatology from ESSA III and V computer produced mosaics. Bull. Amer. Meteor. Soc., 48, 12, 878-883.

Koteswaram, P., 1963: Origin of tropical storms over the Indian Ocean. Proc. Inter-regional Seminar of Tropical Cyclones in Tokyo, 1962. Japan Meteor. Agency, 69-78.

Krishnamurti, T. N. and R. S. Hawkins, 1970: Mid-tropospheric cyclones of the southwest monsoon. J. Appl. Met., 9, 442-458.

________________, 1979: Tropical meteorology, Compendium of meteorology, ed. by A. Wiin-Nielsen. WMO No. 364. 428 pp.

Landsberg, H., 1941: On climatological aids to local weather forecasting. Bull. Amer. Meteor. Soc., 22, 103-105.

La Seur, N. E., 1964: Synoptic models in the tropics. 319-328. Proc. Symposium on Tropical Meteorology. New Zealand Metor. Service, Wellington, New Zealand.

Leipper, D. F., 1967: Observed ocean conditions and Hurricane Hilda, 1964. J. Atmos. Sci., 24, 182-196.

Matsuno, T. A. 1970: Vertical propagation of stationary planetary waves in the winter Northern Hemisphere. J. Atmos. Sci., 27, 871-883.

Miller, F. R. and R. N. Keshavamurthy, 1968: Structure of an Arabian Sea summer monsoon system. I.I.O.E. Meteor. Monograph 1, East-west Center Press, Honolulu.

Murakami, T., 1977: Regional energetics over the North Pacific, South China Sea and the Idonesian Seas during winter. Pageoph, 115, 1283-1301.

Neiuwolt, S., 1977: Tropical climatology. John Wiley & Sons. 207 pp.

Neumann, C. J., G. W. Cry, E. L. Caso, and B. R. Jarvinen, 1978: Tropical cyclones of the North Atlantic ocean, 1871-1977. NOAA, U.S. Dept. of Commerce.

Newell, R. E., J. W. Kidson, D.G. Vincent and G. J. Boer, 1972: The general circulation of the tropical atmosphere and interactions with extratropical latitudes, Vol 1. MIT Press. 258 pp.

Newell, R. E., J. W. Kidson, D.G. Vincent and G. J. Boer, 1974: The general circulation of the tropical atmosphere and interactions with extratropical latitudes, Vol 2. MIT Press. 371 pp.

Oort, A. H. and P. H. Chan, 1977: On the role of the Asian monsoon in the angular momentum and kinetic energy balances of the tropics. Pageoph, 115, 1167-1186.

Ooyama, K., 1963: A dynamical model for the study of tropical cyclone development. Geofisica Internacional, 4, 187-198.

__________, 1982: Conceptual evolution of the theory and modeling of the tropical cyclone. J. Meteor. Soc. Japan, 60, 1, 369-380.

Palmen, E., 1948: On the formation and structure of tropical hurricanes. Geophysica, 3, 26-38.

_________, 1949: Origin and structure of high-level cyclones south of the maximum westerlies. Tellus, 1, 22-31.

_________, 1956: Formation and development of tropical cyclones. Proc. Tropical Cyclone Symp., Brisbane, Australian Bureau of Meteor, Melbourne. 213-231.

_________, and C. W. Newton, 1969: Atmospheric circulation systems, their structure and physical interpretation. Academic Press. 603 pp.

Panosfsky, H. A., 1951: Large-scale vertical velocity and divergence. Compendium of Meteorology. American Meteor. Soc. 1334 pp.

Piersig, W., 1936: Variations of pressure and winds, a treatise on the weather of trade wind region of eastern North Atlantic. Aus dem Archiv d. deutschen Seewarte, Bd. 54, No. 6. English translation of Parts 2 and 3 in Bull. Amer. Meteor. Soc., 25, 2-16, 1944.

Ramage, C. S., 1956: Review of methods of tropical synoptic analysis, with particular reference to tropical cyclones. Proc. Tropical Cyclone Symp, Brisbane, Australian Bureau of Meteor., Melbourne. 213-231.

__________, 1962: The subtropical cyclones. J. Geophys. Res., 67, 4, 1401-1411.

__________, 1974: Monsoonal influences on the annual variation of tropical cyclone development over the Indian and Pacific Oceans. Mon. Wea. Rev., 102, 745-753.

Ramaswamy, C., 1962: Breaks in the Indian summer monsoon as a phenomenon of interaction between the easterly and subtropical westerly jet-streams. Tellus, 14, 337-349.

Rao, Y. P., 1981: The monsoon as reflected in the behaviour of the tropical high-pressure belt. Monsoon dynamics, Lighthill, J and R. P. Pearce, eds. Cambridge Univ. Press. 209-212.

Revell, C. G., 1981: Tropical cyclones in the southwest Pacific. New Zealand Meteor. Service, Wellington. 53 pp.

Riehl, H., 1945: Waves in the easterlies and the polar front in the tropics. Misc. Rep. 17, Dept. Meteor., Univ. Chicago, 79 pp.

_______, 1954: Tropical meteorology. Mcgraw-Hill Book Co., 392 pp.

_______, 1967: Varying structure of waves in the easterlies. Proc. of the International Symposium on Dynamics of Large-scale Process in the Atmosphere. Moscow, 23-30 June 1965. Moscow, 1967, 411-417.

_______, 1979: Climate and weather in the tropics. Academic Press. 611 pp.

Sadler, J. C., 1964: Tropical cyclones of the eastern North Pacific as revealed by TIROS observations. J. App. Met., 3, 347- 366.

__________, 1967a: On the origin of tropical vortices. Proc. of the Working Panel on Tropical Dynamic Meteor. 39-76. Navy Weather Research Facility. NWRF 12-1167-132.

__________, 1967b: The tropical upper tropospheric trough as a secondary source of typhoons and primary source of trade wind disturbances. Final Report, Contract No. AF 19(628)-3860, Hawaii Institute of Geophysics, Report 67-12, 44 pp.

__________, 1976a: Tropical cyclone initiation by the tropical upper tropospheric trough. NEPRF TP No. 2-76. US Navy. 33 pp.

__________, 1976b: A role of the tropical upper tropospheric trough in early season typhoon development. Mon. Wea. Rev., 104, 1266-1278.

__________, B. Kilonsky, L. Oda, and A. Hori: Mean cloudiness over the global tropics from satellite observations. NEPRF Contractor Report 84-09, U.S. Navy. 73 pp.

Shaw, S. L., 1981: A history of tropical cyclones in the central and North Pacific and the Hawaiian Islands 1832-1979. National Weather Service, NOAA, U.S. Dept. of Commerce.

Sikka, D. R., 1977: Some aspects of the life history, structure and movement of monsoon depressions. Pageoph, 115, 1501-1529.

Simpson, R. H., 1952: Evolution of the Kona storms, a subtropical cyclone. J. Met., 9, 24-35.

Tachi, T., 1961: Some statistical studies on tropical cyclones in the western Pacific Ocean. Tenki, 8, 337-344.

Troup, A. J., 1965: The southern oscillation. Quart. J. Royal Meteor. Soc., 91 490-506.

Wallace, J. M., 1973: General circulation of the tropical lower stratosphere. Rev. Geophys. and Space Phys., 11, 2, 191-222.

Wytrtki, K., 1973: Teleconnections in the equatorial Pacific. Science, 180, 66-68.

Yanai, M., 1961: A detailed analysis of typhoon formation. J. Meteor. Soc. Japan, 39, 187-214.

________, 1968: Evolution of a tropical disturbance in the Caribbean Sea region. J. Meteor. Soc. Japan, 46, 86-109.

------------, and Ts. Nitta, 1967: Computation of vertical motion and vorticity budget in a Caribbean easterly wave. J. Meteor. Soc. Japan, 45, 444-466.

________, T, Maruyama, T. Nitta and Y. Hayashi, 1968: Power spectra of large-scale disturbances over the tropical Pacific. J. Meteor. Soc. Japan, 46, 308-323.

Zillman, J. W., ed., 1977: Australian Tropical Cyclone Forecasting Manual. Bureau of Meteor. Dept. of Sci., Australia. 274 pp.

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One of the keys to safe and successful naval operations in the tropics is a thorough understanding of tropical meteorology. The Tropical Cyclone Forecasters Reference Guide is designed primarily as a ready reference for mid-latitude forecasters required to provide tropical meteorology support to staff commanders. This technical note provides a comprehensive overview of tropical climatology and is chapter 2 of the reference guide. Subjects discussed include major factors which affect the tropical climate, tropical synoptic models, wave disturbances, tropical vortices and tropical cyclones.

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